The identification of former terrestrial ice stream dynamics from geomorphic evidence and till architecture: A case study of southwestern Saskatchewan

NORRIS, SOPHIE,LOUISE

The identification of former terrestrial ice stream dynamics from geomorphic evidence and till architecture: A case study of southwestern Saskatchewan

Sophie Louise Norris

Department of Geography

University of Durham

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Declaration

I confirm that no part of the material presented in this thesis has previously been submitted for a degree in this or any other university. In all cases the word of others, where relevant, has been fully acknowledged.

Sophie Louise Norris

Durham University

July 2016

Cover image: Shuttle Radar Topographic Mission imagery showing regional patterns of glacial landforms in Saskatchewan, Canada.

Abstract

A multidimensional study, utilising geomorphological and sedimentological techniques, is conducted to investigate the former dynamics and regional till architecture of terrestrial ice streams during the last (Late Wisconsinan) deglaciation of the Laurentide Ice Sheet. Detailed mapping over a 57,400 km^2 area of southwestern Saskatchewan reaffirmed previous proposals of a southwest trending ice stream demarcated by a corridor of megaflutes and mega-scale glacial lineations (Ice Stream 1). Extending from the Canadian shield to southwestern Saskatchewan this ice stream is cross cut by three (one previously unrecognised) southeast trending ice streams (Ice Streams 2A, B and C). Analysis of the lithologic and geophysical characteristics of 197 borehole samples within these corridors reveals a superimposed till and associated deposits comprising 17 stratigraphic units.

A 3D stratigraphic model of the 57,400 km^2 swath was constructed, by extrapolating data away from boreholes using a nearest-neighbour approach. Using this model the thickness, extent and distribution of these stratigraphic units was delineated allowing the depositional history of the region to be reconstructed and thus the extent of till emplaced during ice stream operation through time and space to be inferred. Reconciling this regional till architecture with the surficial geomorphology reveals that surficial units are spatially consistent with a dynamic switch in flow direction recorded by the cross cutting corridors of Ice Streams 1, 2A, B and C. Thin tills at the centre of the trunk zone of Ice Stream 1 in many places lie unconformably over stratified sediments. This suggests widespread basal sliding may have been subordinate to subglacial sediment deformation but the general thickening of tills towards the lobate terminal margins is consistent with subglacial deformation theory. In addition, variations in till thickness are also recognised on a more localised scale. These variations are attributed to three processes; **1.** down-ice thickening associated with buried valley margins; **2.** upland thinning; and **3.** thickening as a result of overridden glacial-marginal landforms. The significance of newly interpreted patterns of till deposition resulting from ice streaming are then considered and a model of ice stream till deposition is presented. This model provides a generalised view of the pattern of deposition resulting from fast flow over a unlithified sediment bed which may be used to infer the dynamic behaviour of other former ice sheets from their sediment imprint.

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I would firstly like to thank my supervisors Professor David Evans and Professor Colm Ó Cofaigh for their invaluable guidance and support throughout this program, and particularly for their idea of creating 3D models of stratigraphy. Thanks must additionally go to Saskatchewan Research Council for supplying borehole logs, without which this project would not have been possible. I would also like to thank my family, whose assistance and continued support over the past year has been essential in completing this thesis. Last but certainly by no means least I like to thank my friends (particular my Norwegian adventuring, snap playing, waffle eating ones) who have had to put up with my continual chatter about all things ice stream, sediment, and RockWorks related!

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1. Introduction

1.1 Introduction and rationale

Ice streams are fast flowing, highly dynamic corridors within ice sheets capable of obtaining high velocities (12 km a^{-1}) (Alley and Bindschadler, 2001; Clark, 2003) and are of considerable importance due to their dominant role in regulating ice-sheet mass balance (Boulton et al., 1985; Clark, 1992). Though ice streams typically occupy the same flow path, particularly those that are topographically constrained (Vaughan et al., 2001; Lowe and Anderson, 2002), contemporary glaciological research has demonstrated that marine-based ice streams are capable of large changes in flow configuration causing major shifts in flow trajectories over relatively short time scales (10^2 -10^3 yrs) (Jacobel et al., 1996; Conway et al., 2002; Hulbe and Fahnestock, 2004; Dowdeswell et al., 2006). The susceptibility of marine-based ice streams to rapid dynamic flow changes may be a consequence of their being characteristically underlain by saturated, fine-grained, deformable substrates (Thomas, 1979; Dyke and Morris, 1988; Dyke et al., 1992).

Recent work, focused on ancient terrestrial ice streams associated with the Late Wisconsinan, Laurentide Ice Sheet (LIS) deglaciation of the western Canadian Prairies, has revealed that major flow reorganisations are not limited to marine terminating ice streams but also characterise their terrestrial counterparts. Mapping of the distribution of streamlined glacial landforms within southwestern Saskatchewan, by Ross et al. (2009) and Ó Cofaigh et al. (2010), suggests this reorganisation involved a 90° shift in flow direction over a single glaciation. The cause of such a glaciodynamic shift is suggested to result from temporal and spatial variations in the interaction between frozen and thawed bed conditions, with thinning and shutdown of one ice stream triggering initiation of fast flow in another region. This dynamism explains the overprinting (cross cutting) of ice flow indicators (flow-sets) such as drumlins, flutings and mega scale glacial lineations (MSGLs). Although current work in southwestern Saskatchewan has provided a regional scale assessment of surface geomorphology, there is a distinct lack of detailed (local scale) geomorphological mapping and sub-surface sedimentary analysis relating to palaeo-ice stream behaviour. This project will address this limitation by providing the first comprehensive sub-surface stratigraphic and high resolution geomorphic assessment, of a 57,400 km^2 area covering the majority of southwestern Saskatchewan (referred to as the southwestern Saskatchewan swath: see section 2.1), in order to present a more in-depth regional reconstruction and understanding of the spatial and temporal dynamics of palaeo-ice streams in the Interior Plains of western Canada.

1.2 Research design

The research presented here aims to increase current understanding of the relationship between data that are undoubtedly connected to glacial processes (e.g. till architecture) and landform assemblages that are thought to relate to glacial dynamics including ice streaming. This will lead to a more robust subglacial landscape analysis that will improve the way glacial stratigraphic records and geomorphic evidence are analysed and interpreted in the Canadian Prairies. Furthermore this project offers a unique opportunity to assess the till architecture associated with palaeo-ice stream operation and thus test theoretical models established by Boulton (1996a, b) related to till deposition patterns associated with fast glacier flow and ice streaming.

1.2.1 Aims

The overarching aims of this study are to: **i.** reconstruct the depositional history of the southwestern Saskatchewan swath; **ii.** determine the spatial dynamics and regional sedimentary architecture of palaeo-ice streams **iii.** evaluate the significance of depositional patterns in the context of theoretical models of regional till architecture (Boulton 1996a,b).

1.2.2 Research objectives:

To achieve these aims the following set of research objectives were developed:

1. to accurately map the glacial geomorphology of the southwestern Saskatchewan swath from SRTM and Landsat ETM+ imagery.

2. to define Pleistocene stratigraphic units within the southwestern Saskatchewan swath according to lithologic properties, thickness, distribution and genesis.

3. to apply stratigraphic modelling techniques using Rockworks 16™, to construct a regional stratigraphic model.

4. to reconstruct the depositional history of the southwestern Saskatchewan swath and establish the extent of till emplaced during ice stream operation through space and time

5. to critically assess theoretical models established by Boulton (1996a, b) related to till deposition patterns associated with fast glacier flow and ice streaming.

1.3 Thesis outline

The following chapters present the context, location, methodology, results, interpretation, discussion and conclusions of this research. This thesis comprises seven chapters following the introduction. The glacial history of the western Canadian Prairies is reviewed in Chapter 2 with particular attention given to southwestern Saskatchewan. In Chapter 3 ice streams and palaeo-ice streams are examined. This chapter discusses the dynamics and controls associated with contemporary ice streams and the key geomorphological and sedimentological evidence that is used to identify palaeo-ice streams. Within the context of this study, Chapter 4 reviews the use of remote sensing in geomorphological mapping; the methods used to interpret sedimentological data and the applications of Rockworks 16™. The results of this study are contained in Chapter 5. This includes a copy of geomorphological, surficial geology maps, sediment thickness and bedrock topography diagrams, and stratigraphy cross sections of the southwestern Saskatchewan swath (Map Sheets 1-3; see cover insert), as well as detailed descriptions of the geomorphology and sedimentology of the study area. The interpretations of these results are then discussed in Chapter 6. Chapter 7 provides a synthesis of Chapters 5 and 6, drawing links between their context, and discussing the wider implications of this thesis for palaeo-ice stream dynamics. Chapter 8 summarises the major findings of this work and discusses directions for future research.

2. The Glacial History of Southwestern Saskatchewan

As discussed in Chapter 1 subglacial thermo-mechanical conditions favouring fast flow are considered to have been widespread over southwestern Saskatchewan during the last deglaciation (Marshall et al., 1996). This region thus offers a valuable opportunity to advance our understanding of ice stream operation and dynamics. This chapter discusses the study area introduced in section 1.1 and gives an overview of the current understanding of Quaternary stratigraphy, deglaciation models and palaeo-ice stream reconstructions in southwestern Saskatchewan.

2.1 Study area

Saskatchewan is situated in the Canadian Praires and is bordered by the province of Alberta to the west, Manitoba to the east and Northwest Territories to the north (Fig 2.1). Physiographically, this province falls within two regions: the Interior Plains in the south and the Canadian Shield in the north. The rolling landscape of southwestern Saskatchewan is the result of an extended period of fluvial erosion succeeded by multiple Quaternary glaciations and interglaciations which constructed thick sequences of sediment (up to 300 m locally; Klassen, 1989). The 57,400 km^2 study area encompasses the southwestern sector of the Interior Plains (Fig 2.1) ranging from 110°-109° W and 50°-54° N. The area incorporates National Topographic System (NTS) Maps 72K, 72N, 73C and partially covers 73F. As this area does not encompass the whole of southwestern Saskatchewan, the study area is hereafter referred to as the southwestern Saskatchewan swath (SWSS).

2.2 Geology: bedrock and buried valleys

The study area is underlain by crystalline Precambrian rock, which is overlain by a sedimentary rock cover, the Western Canada Sedimentary Basin (Cumming et al., 2012), which thickens to the southwest of the study area, forming a sedimentary wedge ~6000 m near the Alberta foothills (Mossop and Shetsen, 1994). The preglacial landscape was dominated by rivers flowing in large valleys that drained to the north and northeast (Cummings et al., 2012) (Fig 2.2). These buried valleys were eroded into the bedrock and infilled with sediment ranging in age from Tertiary/Early Quaternary (Empress Group) to Wisconsinan time (Stalker, 1968;

Whitaker and Christiansen, 1972; Evans and Campbell, 1995). The generalised distribution of these valleys was first identified at a regional scale on the Prairies over 100 years ago using exposures through channels in Alberta and Montana (Bell 1884; McConnell, 1885; Tyrrell, 1887; Calhoun, 1906; Alden and Stebinger, 1913), and later mapped systematically at a provincial scale in Alberta, North Dakota, and Manitoba using water-wells and boreholes (Paulson, 1982; Betcher et al., 2005) (Fig 2.2). These subsurface studies demonstrated that Prairie buried valleys were commonly much larger than the valleys on the modern land surface. However water-well and borehole based work was more local in perspective and southwestern Saskatchewan remains one of the few areas where buried valleys have not been mapped systematically.

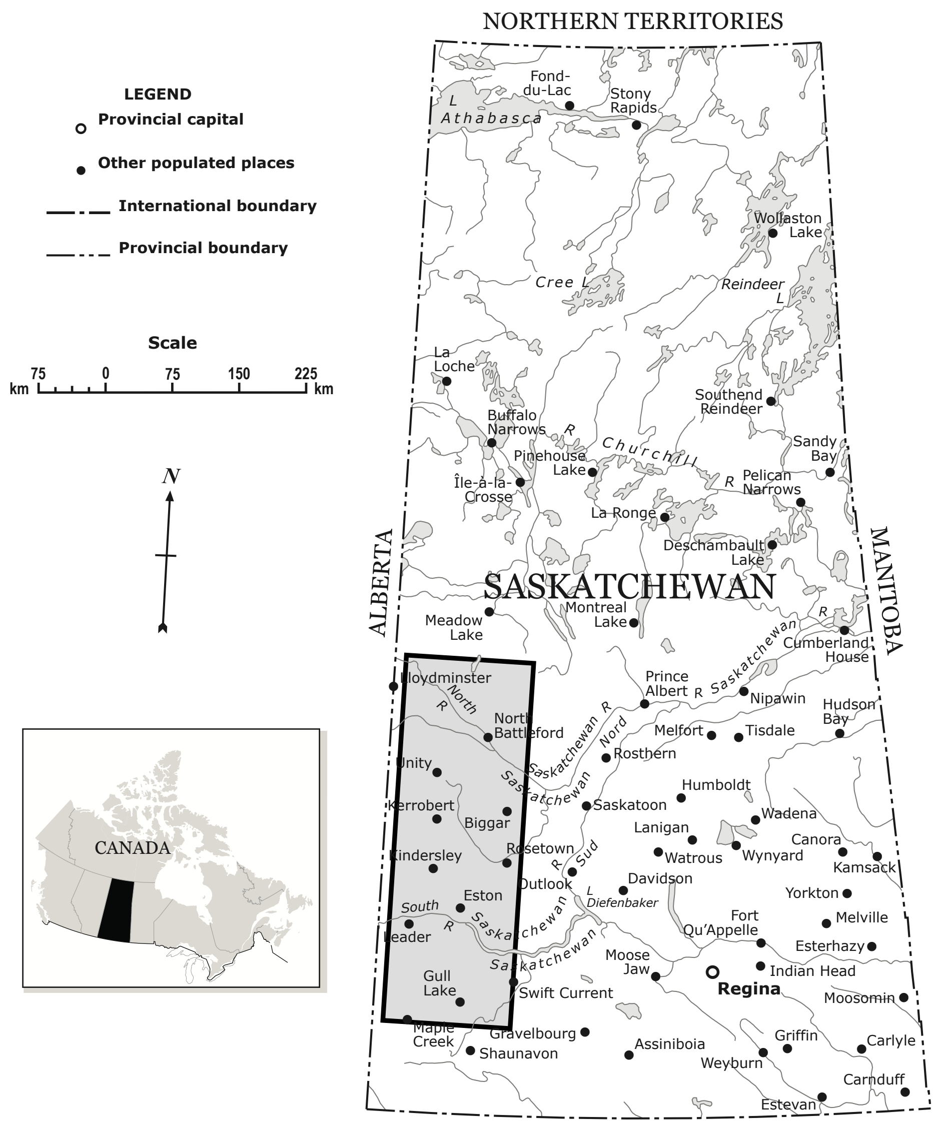


Figure 2.1: *Provincial map of Saskatchewan showing the study area outlined in grey. Adapted from the Atlas of Canada-available at: http://atlas.nrcan.gc.ca/site.html.*

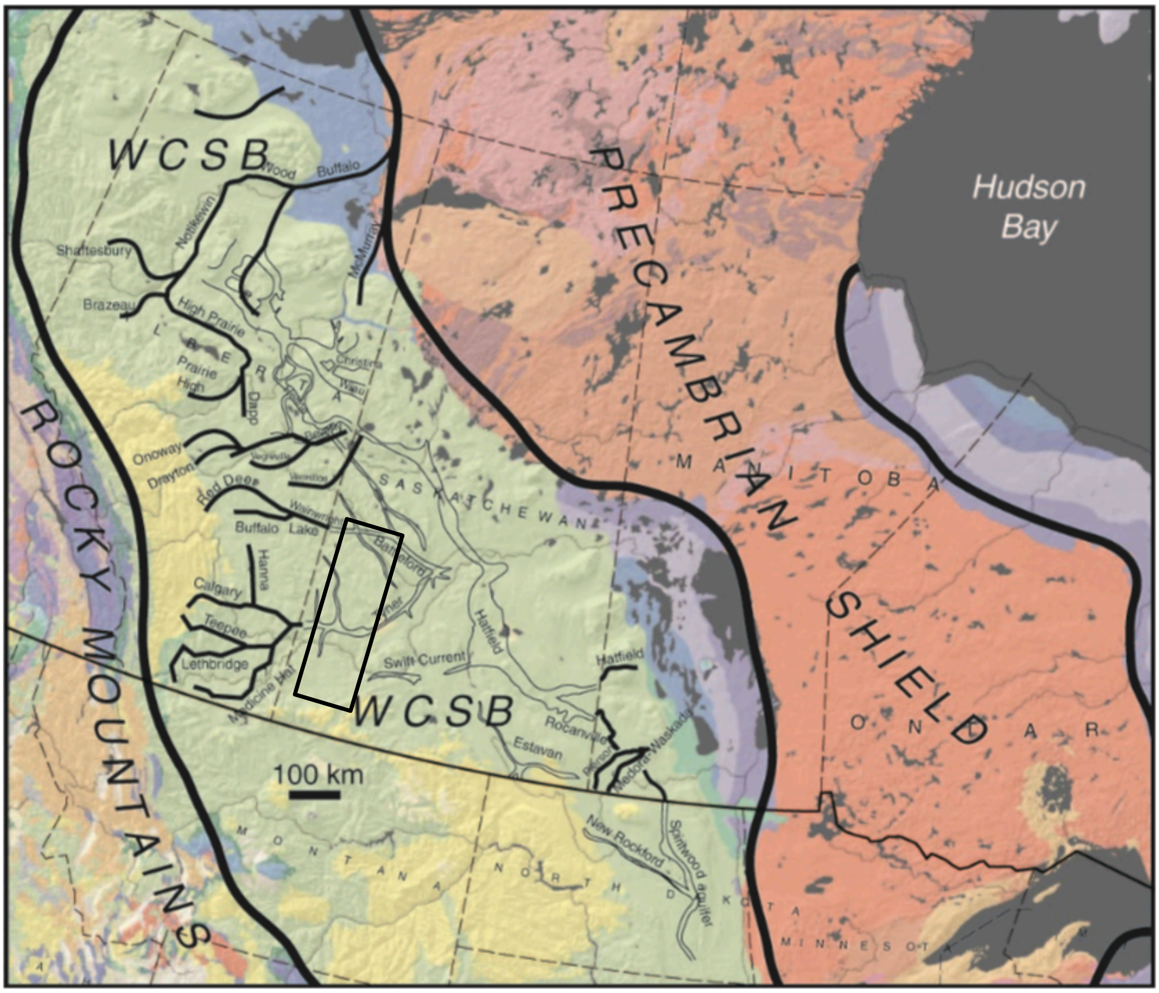


Figure 2.2: *Buried valleys on the Canadian Prairies and North Dakota (WCSB stands for Western Canadian Sedimentary Basin). Buried valley extents are approximately denoted by double grey lines, whereas buried valley thalwegs are indicated by black lines. Red/orange colours indicate Precambrian Shield rocks; blue and purple are Palaeozoic carbonate rocks; green is Mesozoic sedimentary rock (primarily shale); and yellow is Tertiary sedimentary rock. Compiled from: Stalker (1961), Whitaker and Christiansen (1972), Maathuis and Thorleifson (2000) and Oldenborger et al. (2010).*

2.3 Stratigraphy

2.3.1 Late Tertiary to early Quaternary stratigraphy

Sediments of supposed late Tertiary to early Quaternary age were first reported within southern Saskatchewan by McConnell (1885), who identified 'pebble conglomerates and incoherent gravels and silty beds which are found, as valley and lake deposits'. The distribution of these sediments was subsequently assessed on a provincial scale by Rutherford (1937) and Stalker (1968). Since their identification it has become common in Saskatchewan to refer to all 'stratified gravel, sand, silt and clay of fluvial, lacustrine and colluvial origin that overlie bedrock and underlie glacial till of Quaternary age as the 'Empress Group''(Whitaker and Christiansen, 1972). The name is derived from the town of Empress and the type section for the formation is along the east bank of the South Saskatchewan River (Fig 2.3). Drill records indicate three lithologically distinct units can be mapped within the group: unit 1, a basal sand and gravel containing clasts derived only from the Rocky Mountain Cordillera and local bedrock: unit 2, a middle silt and clay, with minor sand and gravels beds: and unit 3, an upper glacial sand and gravel with clasts derived from the Canadian Shield (Andriashek and Fenton, 1989).

2.3.2 Quaternary stratigraphy

The Quaternary sediments covering southwestern Saskatchewan are of glacial, fluvial, lacustrine, organic and aeolian origin. The majority of this sediment is till: sediment transported and deposited by glacier ice with little or no fluvial sorting (Cummings et al., 2012). The characteristics and distribution of tills within the Interior Plains are a reflection of bedrock topography, patterns of ice flow, and the glacial processes that were active during the various phases of glaciation. These tills have been differentiated and correlated within southern Saskatchewan on the basis of their carbonate content, weathering zones, geophysical and geotechnical properties (Christiansen, 1968a; Christiansen 1971; Sauer and Christiansen, 1991). The tills identified in a large number of boreholes and a few stratigraphic sections have been analysed using these criteria leading to the development of a 'layer cake' stratigraphic framework. The framework was first established by Christiansen (1968b) and was later refined (Fig 2.3) (Whitaker and Christiansen 1972; Christiansen 1992; Barendregt et al., 1998). Christiansen (1968a, b) divided the Quaternary deposits in southern Saskatchewan into the lower Sutherland and upper Saskatoon groups and proposed the name Battleford Formation for the surficial till. The Floral Formation was defined as the till lying between the Sutherland Group and the Battleford Formation, while the Sutherland Group (composed of the Warman, Dunburn and Mennon Formations) was defined as the part of the drift lying between the Empress Formation and the Saskatoon Group. The upper till of the Floral Formation has been assigned to the Early Wisconsinan. In contrast the Battleford Formation represents Late Wisconsinan tills. Chronological constraints on the ages of these units are lacking, with the Battleford Formation being the oldest dated unit. The only chronological control on sediments is provided by two sets of radiocarbon ages of 24,500-28,600 ± 560 yrs BP (wood) to 18,000 ± 450 yrs BP (carbonaceous silt) from the stratified layer that underlies the Battleford Formation (Christiansen, 1971).

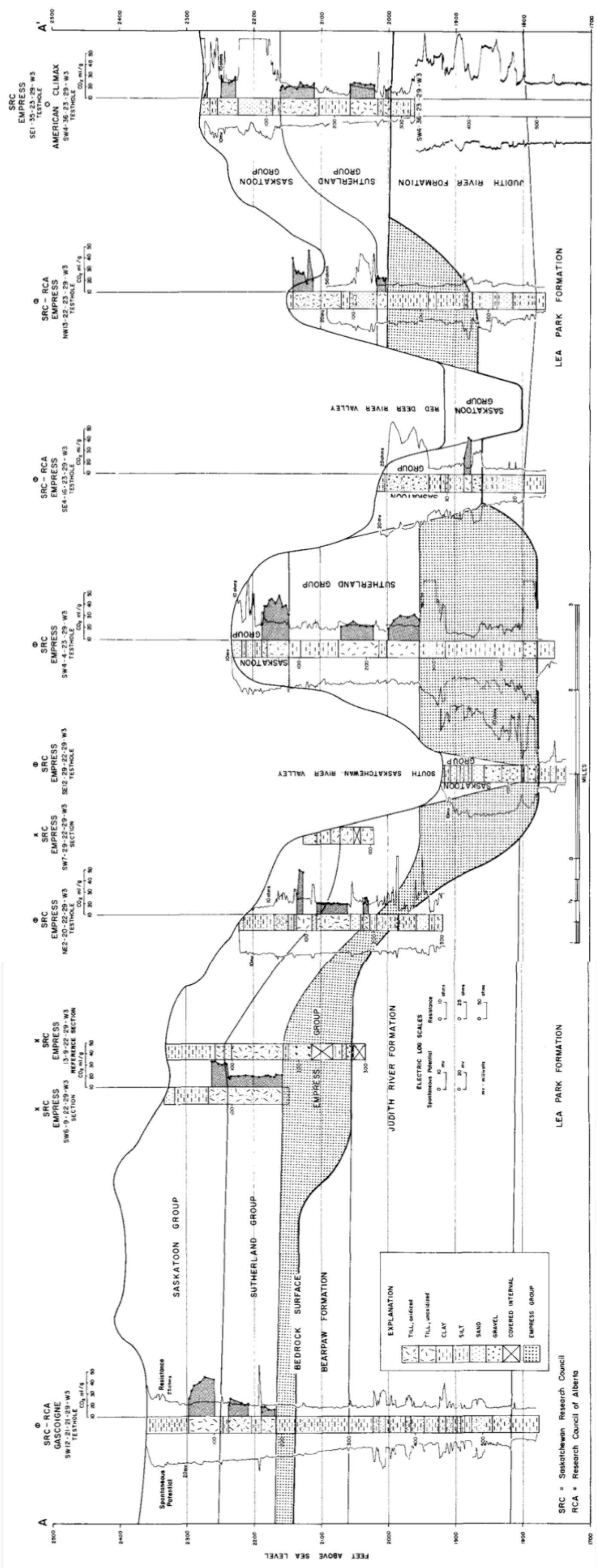


Figure 2.3: *Geological reference section taken from the Empress Area. In the reference section a lower unit of 8 ft (2.44 m) of chert and quartzite gravel and an upper unit of 95 ft (29 m) of interbedded gravel, sand, silt, and clay comprise the Empress Group. Taken from: Whitaker and Christiansen (1972).*

The Battleford Formation is generally thin (Christiansen, 1968a), less than a metre to a few metres, and consists of two till units. Using till consolidation characteristics, the upper unit which is characterised by normal consolidation values has been interpreted as a supraglacial melt-out till, whereas the lower till unit, because of its moderate over consolidation values, has been interpreted as basal melt-out till (Sauer and Christiansen, 1991). The matrix of this subglacial till consists of ~45% sand, 35% silt and 20% clay. The over consolidated state of the pre-Battleford sediments has been interpreted as the result of repeated glacial loading (Sauer et al., 1993). The upper till of the Floral Formation comprises ~45% sand, 30% silt and 25% clay, while the older tills contain over 30% clay (Christiansen, 1992). A high proportion of the clay consists of montmorillonite derived from the bentonitic shales (Scott, 1976). As a result, the tills are characterised by low matrix permeability (Van der Kamp, 2001). This, coupled with evidence that the southern lobes of the LIS exhibited low surface gradients (Mathews, 1974), has prompted many researchers to link ice-sheet instabilities (surging behaviour) in this area to high subglacial water pressures and low bed strength (Clark, 1994) or ice-bed decoupling (Marshall et al., 1996). However no comprehensive sub-surface stratigraphic assessment of these tills in relation to ice stream activity during the Late Wisconsinan deglaciation has yet been undertaken.

Quaternary lithostratigraphic units composed of till can be correlated using outcrop and subsurface data within individual regions on the basis of their composition, geophysical log signature and stratigraphic position. Interprovincial correlation is accomplished primarily using data obtained from widely scattered key outcrops, testholes and waterwells. The chronological data (absolute and relative) for correlations come from a combination of palaeontological, palaeomagnetic, palaeoecological, and tephra studies, together with the stratigraphic position of the units. There is very limited absolute dating within southern Saskatchewan itself (see Section 2.3.2). Subsurface correlation focuses primarily on the composition (obtained from cuttings, core and auger samples) and geophysical log signatures of the till units. Geophysical logs are used extensively in the correlation of Quaternary units. They are used primarily to differentiate till units from stratified units. Thus regional correlations for the entire Interior Plains are tentative. Table 2.1 displays regional correlations made by Andriashek and Fenton (1989), however as stressed in the original text these correlations are highly speculative.

Table 2.1: *Proposed regional correlations and estimated ages (Christiansen, 1972) of lithostratigraphic units compiled. This includes the stratigraphic framework constructed for southern Saskatchewan after Christiansen (1992) and Barendregt et al. (1998). Compiled from: Klassen (1979, 1989), Fulton et al. (1986), Andriashek and Fenton (1989), Christiansen (1992) and Barendregt et al. (1998).*

TIME UNITS			ENVIRON-MENT	STRATIGRAPHIC UNITS					
				Saskatoon Area, Saskatchewan (Christiansen, 1992; and Barendregt et al., 1998)			Sand River Area 73L, Alberta (Andriashek and Fenton, 1989)	Southern Alberta (Fulton et al., 1986; Klassen 1989)	Southern Manitoba (Klassen 1979, 1989)
Quaternary	Holocene		Post-glacial Deglacial	Saskatoon Group	Stratified Drift		Deglacial and Postglacial Deposits	Surficial Deposits	Surface Deposits
Quaternary	Late Pleistocene	Late Wisconsinan	Glacial	Saskatoon Group	Battleford Fm.	Battleford Till (Upper) Battleford Till (Lower)	Grand Central Fm.	Buffalo Lake Till	Arran Fm, Zelena Fm. Lennard Fm.
Quaternary	Late Pleistocene	Late Wisconsinan Middle Wisconsinan	Proglacial Nonglacial Proglacial	Saskatoon Group	Battleford Fm. Floral Fm.	Weathered Zone	Sand River Fm.	Evil-smelling Band	Unnamed Deposits
Quaternary	Late Pleistocene	Early Wisconsinan	Glacial	Saskatoon Group	Floral Fm.	Floral Upper Till	Marie Creek Fm. (Unit 2)	Cameron Ranch Fm.	Minnedosa Fm
Quaternary	Late Pleistocene Early and Middle Pleistocene	Early Wisconsinan Sangamon Illinoian	Proglacial Nonglacial Proglacial	Saskatoon Group	Floral Fm.	Riddel Member		Mitchell Bluff Fm.	Roaring River Clay
Quaternary	Early and Middle Pleistocene	Illinoian	Glacial	Saskatoon Group	Floral Fm.	Floral Lower Till	Marie Creek Fm. (Unit 1)	? ↑ Brocket Till Maunsell Till Labuma Till ↓ ?	? ↑ Shell Fm. Tee Lakes Fm. Largs Fm. ↓ ?
Quaternary	Early and Middle Pleistocene	Illinoian Pre-Illinoian	Proglacial Nonglacial Proglacial	Saskatoon Group Sutherland Group	Floral Fm. Warman Fm.	Weathered Zone	Ethel Lake Fm.		
Quaternary	Early and Middle Pleistocene	Pre-Illinoian	Glacial	Sutherland Group	Warman Fm.	Warman Till	Bonnyville Fm. (Unit 2)		
Quaternary	Early and Middle Pleistocene	Pre-Illinoian	Proglacial Nonglacial Proglacial	Sutherland Group	Dunburn Fm.	Weathered Zone			
Quaternary	Early and Middle Pleistocene	Pre-Illinoian	Glacial	Sutherland Group	Dunburn Fm.	Dunburn Till	Bonnyville Fm. (Unit 1)		
Quaternary	Early and Middle Pleistocene	Pre-Illinoian	Proglacial Nonglacial Proglacial	Sutherland Group	Mennon Fm.	Weathered Zone	Muriel Lake Fm.		
Quaternary	Early and Middle Pleistocene	Pre-Illinoian	Glacial	Sutherland Group	Mennon Fm.	Mennon Till	Bronson Lake Fm.		
Quaternary	Early and Middle Pleistocene	Pre-Illinoian	Proglacial	Empress Group			Empress Unit 3		
Tertiary	Pliocene		Preglacial	Empress Group			Empress Unit 2 Empress Unit 1		

2.3.3 Stratigraphic history: a summary

Based on the stratigraphic sequences described above, multiple attempts have been made to summarise the glacial history of the region (Foster and Stalker, 1976; Fenton, 1984; Klassen, 1989). The key stages are summarised by Fenton (1984) who proposed a two-stage (preglacial, glacial/interglacial) evolution of the region. Prior to glaciation the Interior Plains consisted of broad, southeast-trending valleys separated by low uplands. Streams flowing through these valleys contain deposits composed of quartz sand and gravel dominated by clasts of resistant quartzite, argillite and chert derived from the Cordillera or reworked, by fluvial or glacial processes from older Tertiary and Cretaceous deposits. The Laurentide glaciers advanced across the plains at least five times (Fenton, 1984; Klassen, 1989). Investigations indicate the earliest advance into southern Saskatchewan, was likely prior to 1.8 Ma (Foster and Stalker, 1976; Fenton, 1984; Klassen, 1989). Each advance and subsequent melting resulted in the deposition of glacial and nonglacial units rich in igneous and metamorphic rock fragments derived from the Precambrian Shield, and carbonate fragments excavated from the adjacent belt of Palaeozoic bedrock.

During each ice advance the northeast drainage was dammed meaning lakes developed in valleys and depressions and drainage was diverted southwards (Christiansen, 1979; Clayton et al., 1985). Ice marginal lakes developed as ice retreated and melting took place downslope and steep-walled valleys were cut where: **1.** meltwater flowed through several lake basins; **2.** flow was channeled southward along ice margins; and **3.** drainage was re-established in sediment-filled sections of preglacial valleys. In many places during nonglacial times, the drainage commonly followed the preglacial valleys with channels cutting down into the thick sequence of sediment left behind by retreating ice. Locally, flow was diverted from one valley system to another through trenches excavated by meltwater. Stream deposits laid down during nonglacial periods partly consist of sands and gravels containing resistant pebbles similar to those deposited during preglacial times, but also incorporate distinctive materials from the Precambrian Shield and the adjacent fringe of Palaeozoic carbonate bedrock, transported by westward and southward flowing ice (Fenton, 1984; Klassen, 1989). Repeated glacial and nonglacial intervals left a complex sequence of glacial, fluvial and lacustrine sediment of different ages. While a very generalised provincial wide picture of the regional depositional history is known, little is known about the exact dynamics of these ice sheets, their flow directions and details of the interglacial periods between them. Thus the detailed subsurface investigation provided in this study affords an ideal opportunity to advance current understanding of the glacial evolution of the region.

2.4 Late Wisconsinan deglaciation history

The landforms within Saskatchewan display the prominent geomorphological imprint of the final and most recent (Late Wisconsinan) glaciation, when ice moved eastwards from the Rocky Mountain Cordilleran Ice Sheet and westwards from the LIS (see Klassen, 1989 for review). Coalescence of these two ice masses in the High Plains of Alberta and Saskatchewan resulted in the deflection of ice flow in a southerly direction. Mapping of the glacial geomorphology of southwestern Saskatchewan (Ross et al., 2009; Ó Cofaigh et al., 2010) has enabled a broad identification of flow patterns and landform assemblages, facilitating the development of several Late Wisconsinan deglaciation models. However, the mapping compiled so far lacks the resolution necessary for detailed assessment of regional palaeo-ice stream reconstruction. Furthermore no regional correlation of the subsurface stratigraphy with the complex surface geomorphology has been undertaken.

<u>2.4.1 Late Wisconsinan deglaciation models and palaeo-ice stream dynamics</u>

Christiansen (1979) proposed the first model focused on the Late Wisconsinan deglaciation of southwestern Saskatchewan, which was later expanded to the entire province by the Saskatchewan Geological Survey (2003). Other models encompassing a larger area and using different data sets were published subsequent to this (Dyke and Prest, 1987; Boulton and Clark, 1990; Margold et al., 2015a; Stokes et al., 2016). Most of these models attempt to reconstruct the patterns and chronology of ice retreat, but some also discuss the glacial dynamics, with a major change in ice-flow patterns over the southern Interior Plains being the key component (Dyke and Prest, 1987; Margold et al., 2015a; Stokes et al., 2016). This is particularly reinforced in recent work by Stokes et al. (2016) who infer that the underlying geology and topography of ice streams clearly influenced ice stream activity, but at a ice sheet scale ice streams drainage networks adjusted to changes in ice sheet volume. Based on the recent reconstruction and chronology of the LIS marginal retreat history from Dyke et al. (2003), the regional ice flow was toward the southwest at ~18,000 ^{14}C yrs BP and gradually shifted southward at about 15,000 ^{14}C yrs BP and then southeastward by 13,000 ^{14}C yrs BP. Using Fisher et al.'s (1985) work, Dyke and Prest (1987) briefly discuss the evidence for ice-streaming in different parts of the LIS and propose the late glacial flow shift was driven by soft deforming beds. Broadly, they argue that changes in the basal thermal regime 'made the glacier bed deformable' resulting in the ice-flow patterns becoming more sensitive to regional slope. Clayton et al. (1985) focussed on surging events and their importance to the glacial dynamics of the southwestern LIS, concluding that the fast-flowing ice was caused by a subglacial meltwater system and was thus constrained to areas containing substrates of low permeability. Furthermore, Aber (1993) suggested that successive southeastward surges of Weyman ice lobes ~13,000 to 11,000 ^{14}C yrs BP produced the glacially thrust hills in southern Saskatchewan. Despite these new ideas, regional ice-flow remained loosely constrained and the flow shift remained based on a uniform ice-flow model (Klassen, 1989).

In an attempt to address this issue Ross et al. (2009) and Ó Cofaigh et al. (2010) assessed complex geomorphology in southwestern Saskatchewan and grouped coherent patterns of landforms to reveal two cross cutting palaeo-ice stream landsystems. These ice streams area also mapped as part of large-scale review of LIS ice streams (Margold et al., 2015a, b). One palaeo-ice stream flowed southwest (Ice Stream 1) while the other flowed southeast (Ice Streams 2a, b) (Fig 2.4). It should be noted that Ice Stream 1 and Ice Streams 2 (a, b) have also been previously referred to as the Maskwa and Buffalo ice stream systems respectively (Ross et al., 2009), though are referred to Ice Stream 1 and 2 hereafter. Ice Stream 1 is represented, in Fig 2.4, by discontinuous geomorphology in a corridor, which crosses pre-glacial valleys and topographic highs with little apparent disturbance to its operation. The ice stream was by definition relatively thick and basal shear stresses high. In contrast, Ice Stream 2 was thinner and topographically constrained. Importantly Ice Stream 2 fed into the James Lobe and was therefore potentially crucial to the development and sustenance of the southern lobes of the LIS (Ross et al., 2009).

Ross et al. (2009) and Ó Cofaigh et al. (2010) also suggest, based on the chronology outlined by Dyke et al. (2003), that the glacio-dynamic shift from Ice Stream 1 to Ice Stream 2 occurred between ~13,500 yrs BP (Fig 2.5b) and after the last major surge of the James Lobe ~12,500 yrs BP (Dyke et al., 2003). Ice Stream 2 then evolved into thin outlet lobes until final deglaciation of the area at ~10,000 yrs BP (Fig 2.5c). This suggests that, unlike the ice streams found elsewhere in the LIS, the ice streams and outlet glaciers of the southern lobes terminated in terrestrial settings and underwent continual re-arrangement (both advance and retreat) during overall deglaciation. In the LIS, it therefore appears that the terrestrially terminating ice streams along the southern margin of the LIS responded much more rapidly to climate or atmospheric temperature changes, particularly during the early stages of deglaciation (Evans et al., 2008). Southwestern Saskatchewan therefore contains evidence for at least two generations of ice streams that were active during the last deglaciation. The underlying cause of this change in flow configuration is still uncertain, however Ross et al. (2009) and Ó Cofaigh et al. (2010) propose that variation in the relative timing of ice stream retreat or shutdown with reduced ice thickness and other controlling factors (warming trend, variation in subglacial conditions) brought the Ice Stream 1 to a threshold resulting in 90° shift in flow. While both of these studies propose generalised views of the palaeo-ice stream dynamics in the region, due to the low resolution of geomorphological mapping and the lack of any analysis of the subsurface sediment relating to each of these ice streams, there is still a large amount of uncertainty relating to their behaviour and evolution.

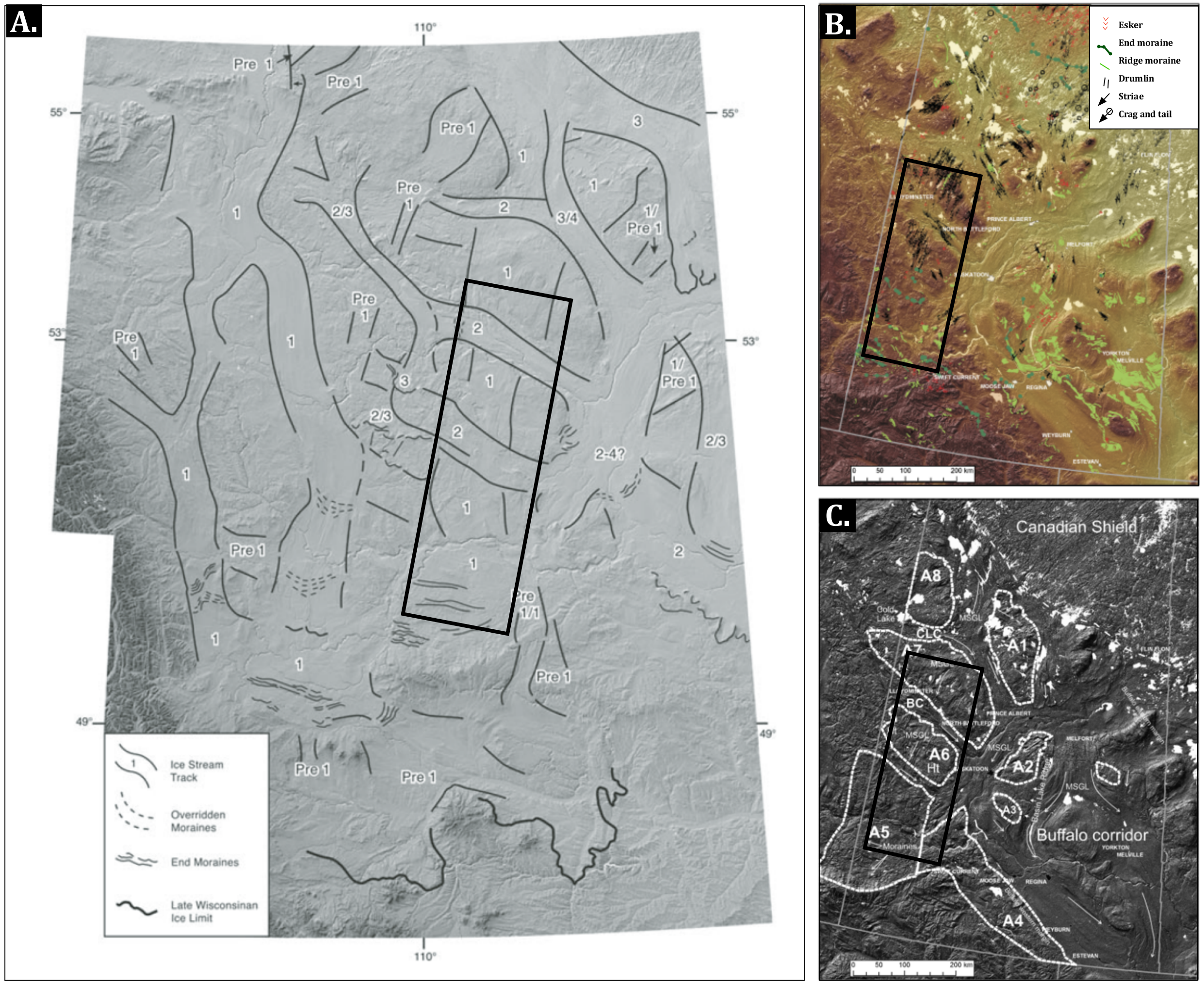


Figure 2.4: *Shuttle Radar Topography Mission (SRTM) derived images of the glacial landscape of southeastern Alberta and southern Saskatchewan. Study area outlined in black.* ***A.*** *Palaeo-ice stream tracks in southeastern Alberta and southern Saskatchewan inferred based on the relationships of crosscutting landforms. Note numbers 1 and 2 within the study area correspond to the Ice Streams 1 and 2 respectively. Taken from: Ó Cofaigh et al. (2010)* ***B.*** *Mapped glacial landforms in southern Saskatchewan. Adapted from: Ross et al. (2009).* ***C.*** *First order sediment landform assemblage overlying a SRTM derived hill shade model. White dashed lines represent the boundary of the ice stream landsystem assemblages correlating with upland areas (labelled A1-8).*

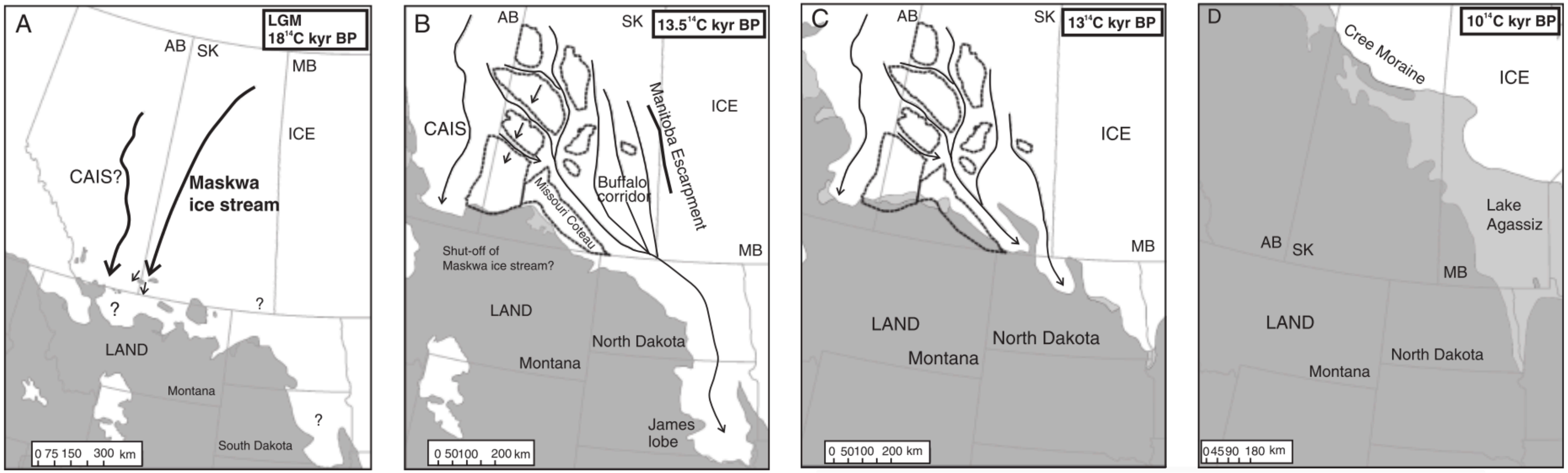


Figure 2.5: *Late Wisconsinan ice stream evolution model showing the shift in ice stream behaviour over time. Taken from: Ross et al. (2009).* ***A.*** *The location of the Maskwa (Ice Stream 1) and Central Alberta paleao-ice stream (CAIS) (see Evans et al., 2008) at the LGM 18,000 ^{14}C yr BP. Both ice streasm trend north south and are not constrained by topography* ***B.*** *Shutdown of the Maskwa palaeo-ice stream (Ice Stream 1) ~13,500 ^{14}C yr BP.* ***C.*** *Late glacial ice stream outlet lobes of the Buffalo corridor (Ice Stream 2) flowing southeast 13,000 ^{14}C yr BP. These ice streams cross the original path of the Maskwa ice stream and are topographically constrained* ***D.*** *Final retreat of the ice sheet onto the Canadian Shield at 10,000 ^{14}C yr BP.*

2.5 Summary

Quaternary sediments overlying southeast-trending valleys and buried channels suggest that the LIS advanced across the plains at least 5 times. The tills deposited have been differentiated and correlated within Saskatchewan on the basis of their composition, geophysical log signature and stratigraphic position. The surface geomorphology displays the prominent imprint of the final and most recent glaciation where a complex subglacial landscape representing two distinct and cross cutting palaeo-ice streams are identified. However mapping compiled so far in places lacks the resolution necessary for the detailed assessment of regional palaeo-ice stream reconstruction. Furthermore no regional correlation between till stratigraphy, ice flow indicators and Late Wisconsinan ice sheet dynamics has yet been established. This location therefore presents an ideal opportunity to investigate the sub-surface stratigraphic architecture in order to present a regional reconstruction and understanding of the spatial and temporal dynamics of ice streams in the Interior Plains of western Canada during the Late Wisconsinan deglaciation.

3. Ice Streams: Dynamics, Controls and Reconstruction

The significance of investigating the landform-sediment record of southwestern Saskatchewan within the framework of the ice stream paradigm was established in Chapter 2. To place this discussion into context, the following chapter will discuss the different types of ice streams; the mechanisms that facilitate fast ice flow; as well as the controls on their operation. Current knowledge of palaeo-ice streams will then be reviewed including characteristic criteria used to identify them in the geomorphological and sedimentological record.

3.1 Ice streams

During the late 1970s the notion that modern day ice sheets contained fast flowing outlets and ice streams became widely acknowledged (Rose, 1979). This idea provided the basis for major research projects in Greenland and Antarctica (e.g. Alley et al., 1986; Bindschadler et al., 1987), which yielded insights into the controls on these highly dynamic sections of ice sheets. Under the 'new' paradigm of ice stream-controlled flow systems (i.e. replacing the concept of uniform flow), it was hypothesised that if contemporary ice sheets contained such outlets then ice sheets that formed during the last glaciation may also have contained similar areas of fast flow (Denton and Hughes, 1981). Thus speculation followed that reconstructions of palaeo-ice sheets must account for such features (e.g. Dyke and Prest, 1987; Boulton and Clark, 1990).

The true extent and locations of palaeo-ice streams has only within the last 20 years started to be fully understood and conceptualised (e.g. Stokes and Clark, 1999; 2001; Clark and Stokes, 2003; Winsborrow et al., 2004; Margold et al., 2015a; Stokes et al., 2016), following the benchmark study of Dyke and Morris (1988) on the geomorphology of palaeo-ice streams. This was facilitated by significant advances in remote sensing techniques and the availability of continental scale data coverage. Terrestrial observations of past fast flow have also been replicated by marine geophysical investigations of the offshore record from glaciated continental shelves in high latitude oceans (e.g. Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002). The coupling of contemporary and palaeo-ice stream data has also helped to strengthen modelling capability and provide strong evidence that, during the LGM, ice streaming occurred across large areas of Europe and North America (e.g. Stokes and Clark, 2001; Clark and Stokes, 2003; Winsborrow et al., 2004; De Angelis and Kleman, 2005; Dowdeswell et al., 2006; Kleman and Glasser, 2007; Margold et al., 2015a; Stokes et al., 2016) and regional reconstructions that incorporate their temporal evolution have also been proposed (De Angelis and Kleman, 2005, 2007; Evans et al., 2008; Ross et al., 2009; Stokes et al., 2009; Ó Cofaigh et al., 2010; Ross et al., 2011).

3.1.1 Ice stream types

Using examples of ice streams in Antarctica and Greenland, Stokes and Clark (1999) proposed that contemporary ice streams could be broadly classified into two categories as either ‘pure’ or ‘topographic’. They highlighted ice streams of the Siple Coast in West Antarctica as good examples of pure ice streams because they do not appear to lie in bedrock troughs (Stokes and Clark, 1999). In contrast, Jakobshavn Isbræ in Greenland is exemplified as a topographically constrained ice stream as its location is largely governed by the underlying bedrock topography, (Stokes and Clark, 1999). The classification of ice stream types was later expanded upon by Truffer and Echelmeyer (2003). The ‘pure’ and ‘topographic’ terms of Stokes and Clark (1999) were replaced with ‘ice-stream’ and ‘isbræ’ types (Truffer and Echelmeyer, 2003). The Siple Coast ice streams were still used as examples of archetypal shallow and soft-bedded ‘ice streams’ and Jakobshavn Isbræ was suggested as an example of a typical ‘isbræ’ (Truffer and Echelmeyer, 2003). Due to the mixture of these four different terms potentially creating confusion, especially as isbrae only covers topographically constrained ice streams in marine environments, this study will refer to ‘pure’ ice streams and ‘topographic’ ice streams, rather than ‘ice-stream’ and ‘isbræ’.

Unlike the earlier identification of ice streams into two separate and distinct types by Stokes and Clark (1999), Truffer and Echelmeyer’s (2003) suggested that these two types should be viewed as opposite ends of a continuum rather than as two distinct entities. This notion is supported by numerous examples of both types of flow in the same ice stream, such as Rutford Ice Stream and Pine Island Glacier in West Antarctica (Truffer and Echelmeyer, 2003) and Lambert Glacier in East Antarctica (Hambrey and Dowdeswell, 1994). Even the ‘pure’ Siple Coast ice streams have topographically controlled tributaries further up-ice (Joughin et al., 1999). It is also important to note that ice streams can be classified as 'marine' or 'terrestrial' depending on the environment in which they terminate (Stokes and Clark, 1999). All contemporary ice streams are marine terminating (Stokes and Clark, 1999). However several examples of ice streams that terminate on land either as lobate forms or that calve into large proglacial lakes have been identified within past ice sheets (Fig 3.1). The importance and characteristics of terrestrial terminating ice streams are discussed in section 3.2.1.

3.1.2 Mechanisms of flow

The continuum of ice stream types discussed above can be linked to the proposed mechanisms for fast ice flow (Truffer and Echelmeyer, 2003). The first mechanism that was suggested involved elevated basal water pressure leading to enhanced basal sliding (Rose, 1979). In this case water at the base of an ice sheet reduces the basal shear stress between ice and substrate, increasing the speed of part of the ice sheet (Bell, 2008). Secondly, the internal mechanics of ice deformation were suggested by Hughes (1977) as being responsible for fast flow. This focused on enhanced basal and marginal-ice shear due to stress-induced recrystallisation within the ice (Hughes, 1977). A third mechanism for fast flow within an ice stream was developed following the discovery of a layer of sediment beneath the Whillans Ice Stream, West Antarctica by Blankenship et al. (1986). This mechanism is based on the Boulton and Jones (1979) deformable bed theory, in which saturated sediments deform. Boulton and Hindmarsh (1987) proposed that this could account for up to 90% of an ice stream's basal motion. However the extent to which subglacial deformation is the dominant mechanism for fast ice flow has been shown to be variable. For example ~90% of the forward movement of Ice Stream D on the Siple Coast has been attributed to subglacial deformation, contrasting with only 25% of the motion of the Whillans Ice Stream (Kamb, 2001; Bennett, 2003). Furthermore later studies also showed that the Whillans Ice Stream was found to be melting at its base, allowing for a combination of basal sliding and subglacial deformation to operate (Engelhardt et al., 1990). This demonstrates that basal hydrology plays an important role in both basal sliding and subglacial deformation (Bennett, 2003) and in the spatial and temporal variation of these two forms of flow within an ice stream (Alley, 1989; Boulton et al., 2001a). These three proposed mechanisms (internal ice deformation, basal sliding and subglacial deformation) for rapid flow in ice streams can be linked to the continuum of ice stream types (Stokes and Clark, 1999; Truffer and Echelmeyer, 2003). Pure ice streams have thinner ice, lower surface slopes, low driving stresses, and motion is mainly thought to be a result of basal sliding and subglacial deformation (Engelhardt et al., 1990; Bennett, 2003). Topographically constrained ice streams are thought to have thicker ice, steeper surface slopes, higher driving stresses and significant internal ice deformation (Lüthi et al., 2002). Additionally, there is a tendency for ice acceleration within topographically constrained corridors (Bennett, 2003). It is possible for basal sliding to occur within topographically controlled ice streams as at Jakobshavn Isbræ (Truffer and Echelmeyer, 2003; Roberts and Long, 2005), and for ice streams to vary down-ice from pure to topographically constrained (e.g. Lambert Glacier, East Antarctica; Hambrey and Dowdeswell, 1994) and *vice-versa* (e.g. Siple Coast Ice Streams; Joughin et al., 1999).

While a plethora of research has been undertaken associated with the mechanisms of ice stream flow, much of this research, especially early studies, focus on the Siple Coast Ice Streams (Jennings, 2006). In contrast to the majority of ice streams in Antarctica these ice streams are unusual in that they do not simply flow though sediment-poor deep troughs (Jennings, 2006). However the Siple Coast Ice Streams are probably the closest modern analogue for terrestrial ice streams flowing over soft sediment in palaeo-ice sheets (Winsborrow et al., 2004; Jennings, 2006), and thereby, in relation to this study, prove a useful analogy, which can aid our understanding of palaeo-ice stream dynamics of the LIS during the Late Wisconsinan.

3.1.3 Controls on ice stream dynamics

Multiple theories have been proposed to explain ice stream shutdown. The shutdown of Ice Stream C on the Siple Coast ~150 yrs BP has been the focus of many of these theories due to it being one of the few preserved examples of ice stream shutdown in an active setting (Bennett, 2003). With reference to Ice Stream C the loss of a lubricating basal sediment layer as a result of basal erosion within the ice stream was initially suggested by Retzlaff and Bentley (1993) as the cause of ice stream shut down. The exhaustion of till is one of the four primary causes of localised patches of basal friction called 'sticky spots' (Stokes et al., 2007). Within an ice stream where subglacial deformation is the dominant mechanism of basal motion, an increase in till-free areas in an ice stream could lead to shutdown through sediment exhaustion (Stokes et al., 2007). A second possible cause of ice stream shutdown is via a reduction in the amount of subglacial water (either as a result of water capture by another ice stream or the collapse of a high pressure drainage network into a channelised system) that can lubricate the ice bed (Bennett, 2003). Reducing the amount of water in the subglacial till layer could produce patches of well-drained stronger till, thought to be another major cause of sticky spots (Stokes et al., 2007). In the same way that water could be captured by a neighbouring ice stream, Payne and Dongelmans (1997) proposed that the accumulation area of ice could also be captured, affecting the accumulation areas and thus flow speed of neighbouring ice streams (Payne and Dongelmans, 1997). Changes in thermal processes are the third proposed cause of ice stream shutdown (Bennett, 2003). This is thought to be through a switch from melting to freezing basal conditions (Christoffersen and Tulaczyk, 2003). Rapid ice flow will cause ice thinning (Bennett, 2003). This in turn will cause the movement of colder ice closer to the bed, leading to a reduction in fast flow (Bennett, 2003). While the factors highlighted have been proposed in the context of contemporary ice streams, temporal and spatial variation of several of these factors (surging, loss of lubricating till, ice piracy, water piracy and changes in thermal processes) have been suggested to have played a significant role in the location and dynamics of palaeo-ice streams (Kleman and Glasser, 2007; Ross et al., 2009 and Ó Cofaigh et al., 2010)

3.2 Palaeo-ice streams

In parallel to modern ice streams, palaeo-ice streams are thought to have performed a significant part in the dynamics, stability and evolution of former ice sheets (De Angelis and Kleman, 2005, 2007; Ó Cofaigh et al., 2008). Thus identification of the beds of palaeo-ice streams are critical for a number of reasons. Firstly, in order to accurately reconstruct former ice sheet dynamics and histories we need to know where and when ice streams operated (Stokes and Clark, 2001, 2003; De Angelis and Kleman, 2007). This was first recognised by Denton and Hughes (1981), who predicted the location of ice streams in their reconstruction of the former northern hemisphere ice sheets (Fig 3.1). Secondly, in ordered to assess the interactions and response of past ice sheets to climate, an understanding of the location and behaviour of palaeo-ice streams is vital. Thirdly, unlike the beds of contemporary ice streams that are extremely difficult to study due to their inaccessibility, the beds of palaeo-ice streams are very accessible and present an excellent opportunity to investigate the processes that occur beneath ice streams (Stokes and Clark, 1999; Stokes, 2002; Ó Cofaigh et al., 2005). Finally, as all contemporary ice streams are marine-terminating, palaeo-ice streams provide a unique chance to also study terrestrially terminating ice streams (Stokes and Clark, 1999).

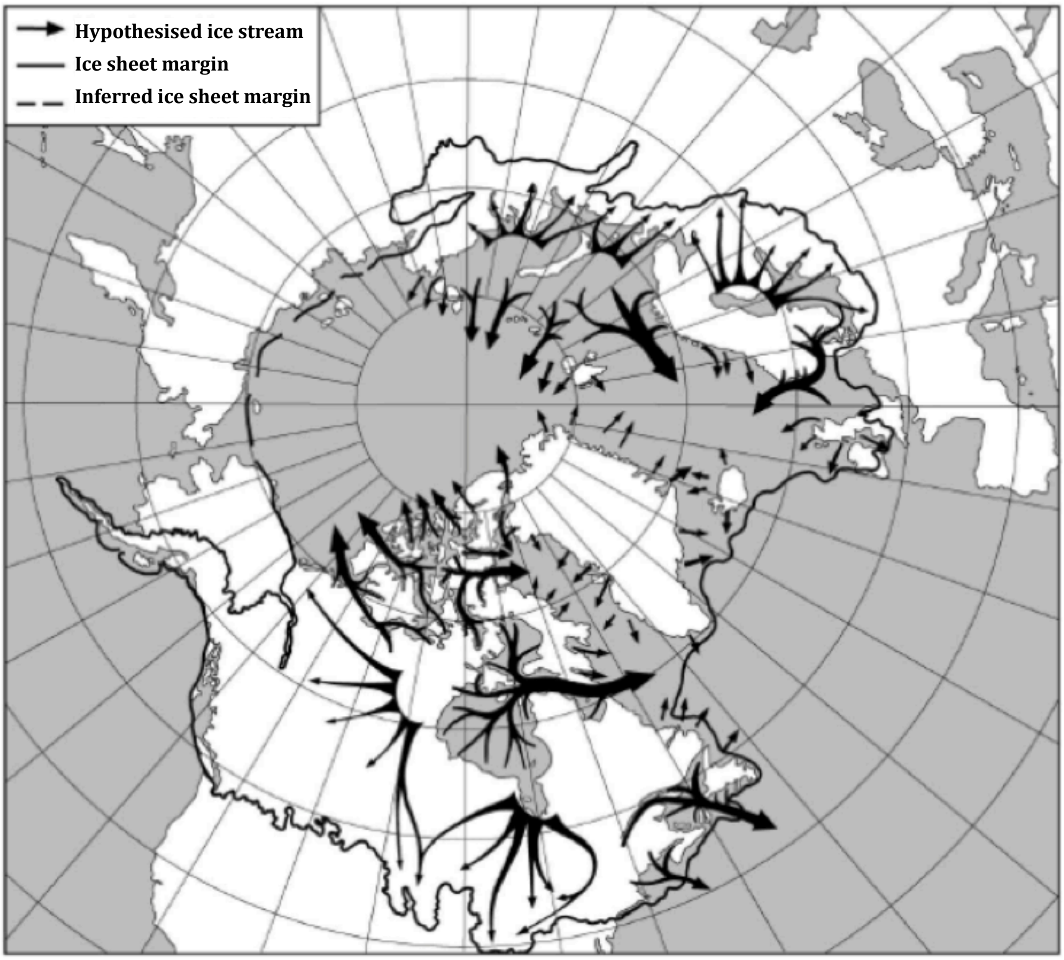


Figure 3.1: *Locations of palaeo-ice stream (indicated by black arrows) in the former northern hemisphere ice sheets as hypothesised by Denton and Hughes (1981).*

3.2.1 Terrestrial palaeo-ice streams

As outlined in section 3.1.1, former ice sheets are thought to have likely drained not only through marine terminating ice streams but also through terrestrial ice streams (Patterson, 1997; Clark and Stokes, 2003). Examples of these ice streams are found at the southern margin of the former LIS (e.g. Patterson, 1997, 1998; Evans et al., 2008, 2012, 2014) as well as the southern sector of the former Fennoscandian Ice Sheet (Boulton et al., 2001b). There are no modern analogues for land-terminating ice streams (Patterson, 1997; Boulton et al., 2001b). Thus, it has been suggested that both modern fast moving or surging glaciers (Patterson, 1997) and piedmont lobe glaciers (Boulton et al., 2001b) present the best contemporary models.

The principal consideration associated with terrestrial ice streams is the way in which ice is removed from their margins (Clark and Stokes, 2003). In the case of marine-based ice streams this is achieved via calving directly into the ocean or from a floating ice shelf (Patterson, 1997; Clark and Stokes, 2003). Terrestrial ice streams are thought to ablate large ice fluxes in two ways; either through calving into a large proglacial lake or through a splayed lobe that extends beyond the ice sheet margin (Patterson, 1997, 1998; Clark and Stokes, 2003; Stokes and Clark, 2004). The large surface area presented by the splayed lobe and its location below the equilibrium line altitude would allow it to facilitate efficient surface melting and mass loss (Clark and Stokes, 2003).

3.2.2 Identifying palaeo-ice streams

A number of approaches guided by the glaciological inversion technique (i.e. methods that formalise the procedure of using the landform record to reconstruct ice-sheet configuration (Kleman and Borgström, 1996)) have been undertaken to investigate both marine and terrestrial palaeo-ice streams, ranging from local to large scale remote sensing projects on a regional or continental scale. However the need to establish a set of criteria by which ice streams could be identified in formerly glaciated terrain was noted by Stokes and Clark (1999). Identifying the geomorphological signatures of ice streams allows the overall landsystem 'fingerprint' (Everest et al., 2005; Golledge and Stoker, 2006) to be compared with signatures of glacial landscapes (Hart, 1999; Stokes and Clark, 2001). Stokes and Clark's (1999) criteria are based on the fundamental characteristics of contemporary ice streams and fast ice flow lobes, however it should be noted that individual criteria may not be exclusive to ice stream activity, but that the collective landsystem can be thought to be characteristic of ice streaming (Stokes and Clark, 1999). The geomorphological criteria for palaeo-ice streaming proposed by Stokes and Clark (1999) are summarised in Table 3.1, and the conceptual signatures of ice streams in marine and terrestrial settings that have deposited landforms isochronously or time transgressively are shown in Figure 3.2.

Table 3.1: *The criteria for the identification of palaeo-ice stream identification as proposed by Stokes and Clark (1999).*

Contemporary ice stream characteristic	Proposed geomorphological signature
Characteristic shape and dimensions	1. *Characteristic shape and dimensions* 2. *Convergent flow patterns*
Rapid velocity	3. *Highly attenuated bedforms* 4. *Boothia-type erratic dispersal trains*
Sharply delineated shear margin	5. *Abrupt lateral margins* 6. *Lateral shear margins*
Deformable bed conditions	7. *Evidence of pervasively deformed till* 8. *Submarine till delta or sediment fan*

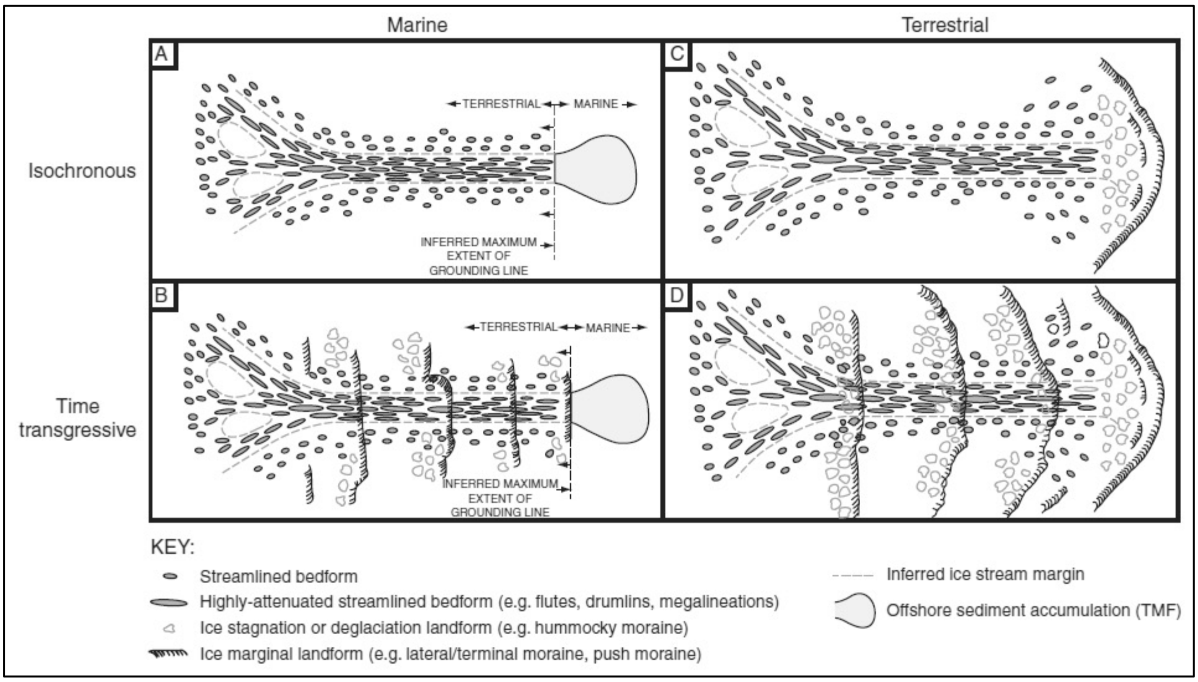


Figure 3.2: *The palaeo-ice stream fingerprint. Taken from: Clark and Stokes (2003). Four end members are recognised:* ***A.*** *marine isochronous;* ***B.*** *marine time transgressive* ***C.*** *terrestrial isochronous; and* ***D.*** *terrestrial time-transgressive. An isochronous or 'rubber-stamped' record may be produced when ice streams stop suddenly or 'switch off' and a largely intact record of the ice streams' activity is preserved in the landscape (Stokes and Clark, 1999, 2001; Jansson et al., 2003). A synchronous record implies that the landforms record true flow patterns.*

3.2.2.1 Characteristic shape and dimensions

The most distinct characteristics of active and palaeo-ice streams are their overall shape and dimension (Stokes and Clark, 1999). The dimensions of ice streams have been analysed by a variety of authors (Stokes and Clark, 1999, 2003; Stokes, 2002; Whillans et al., 1987; Engelhardt et al., 1990; Bennett, 2003; Rignot et al., 2011; Margold et al., 2015a) and are documented to span across a wide range of sizes with lengths from tens to hundreds of kms and widths of hundreds of meters to more than a hundred km (Rignot et al., 2011 Margold et al., 2015a) (Fig 3.3). The Siple Coast Ice Streams that drain West Antarctica are between ~30-50 km wide and ~300-500 km long (Whillans et al., 1987; Engelhardt et al., 1990; Bennett, 2003) and the M'Clintock Channel Palaeo-Ice Stream was reconstructed as 140 km wide and 740 km long (Clark and Stokes, 2001). Additionally Margold et al. (2015a) highlighted the lengths of very long LIS ice streams including the James Lobe, Des Moines Lobe, Hudson Strait and McClure Strait ice streams all of which are reconstructed at >1200 km in length. In contrast much smaller ice streams have also been reported, for example De Angelis and Kleman (2008) identified three palaeo-ice streams in northeast Canada <70 km long, thereby demonstrating the potential variation in ice stream dimensions (Stokes and Clark, 1999).

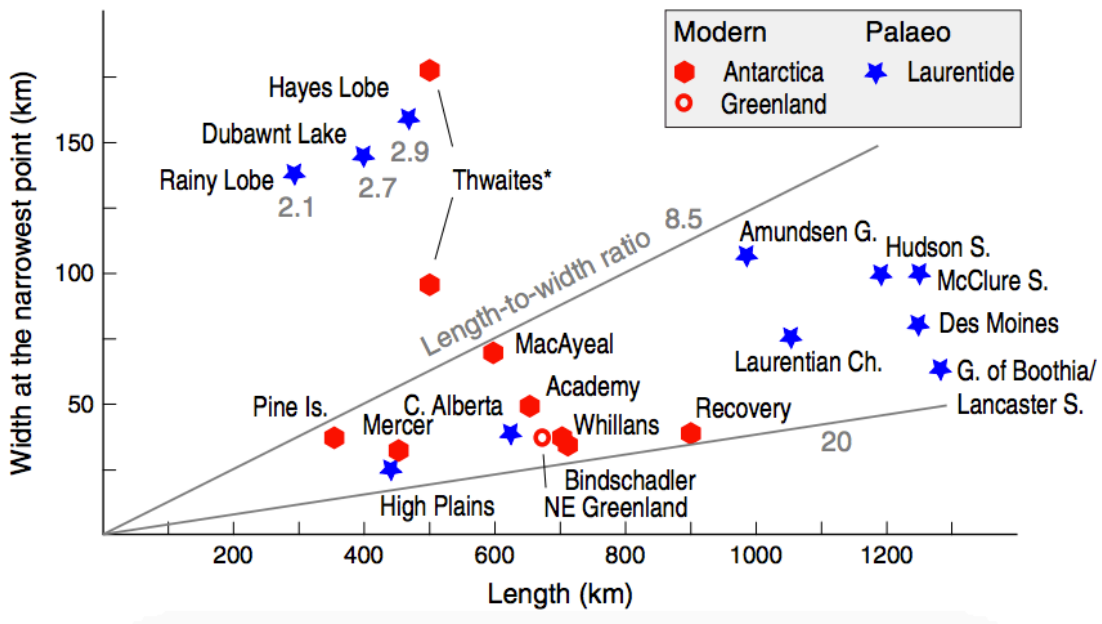


Figure 3.3: *Length-to-width ratios of large ice steams in present-day Antarctic and Greenland ice sheets and in the LIS. Note the Thwaites Glacier is shown twice as it appears to have an outer and inner lateral margin. Taken from: Margold et al. (2015a).*

3.2.2.2 Convergent flow patterns

A convergent pattern of ice flow has been proposed as a key characteristic of ice streams (Stokes and Clark, 1999). The onsets of contemporary ice streams are characterised by large convergence zones, where surrounding slower moving ice is integrated by the ice stream channel (Hodge and Doppelhammer, 1996). For example the Dubawnt Lake Palaeo-Ice Stream (Stokes and Clark, 2003), the Tweed Palaeo-Ice Stream (Everest et al., 2005), the Strathmore Palaeo-Ice Stream (Golledge and Stoker, 2006), and a palaeo-ice stream in Marguerite Bay, offshore of the Antarctic Peninsula (Anderson and Oakes Fretwell, 2008) all show highly convergent flow patterns. However more recent work has also highlighted that not all ice streams will display this convergent onset zone and in contrast some modern ice sheets may form dendritic ice stream networks (Margold et al., 2015a).

3.2.2.3 Highly attenuated bedforms

Linear subglacial bedforms (flutes, drumlins, megaflutes, megadrumlins and MSGLs) aligned parallel to former ice-flow directions are common in areas of formerly glaciated terrain (Clark, 1993, 1994). The largest of these features, MSGLs, were first detected by Clark (1993) and were much larger in scale and had very high length-to-width ratios compared to other streamlined landforms. This ratio is usually of the order of >10:1, and Clark (1994) suggested they could be formed by either fast ice flow over a short duration or slower moving ice over a longer duration.

Building on the work of Clark (1993) it has been suggested that the presence of such elongated features may record fast ice flow within an ice stream (Stokes and Clark, 1999) and the presence of MSGLs and other streamlined bedforms has thus been invoked as evidence for a number of palaeo-ice streams (e.g. Clark and Stokes, 2001; Jansson et al., 2003; De Angelis and Kleman, 2005, 2007; Ó Cofaigh et al., 2002, 2005, 2008; Stokes et al., 2005, 2006, 2009; Golledge and Stoker, 2006; Evans et al., 2008; Ottesen et al., 2008). However the genesis of streamlined bedforms and MSGLs in particular, is not well understood and is the subject of conflicting hypotheses (Clark, 1993; Clark et al., 2003; Shaw et al., 2008). It was suggested by Shaw et al. (2008) that MSGLs are formed through erosion by turbulent meltwater floods. Known as the subglacial megaflood theory of drumlin and attenuated lineation formation, this theory has not been widely accepted and has been criticised by Benn and Evans (2006) who argued that it is incompatible with a large body of mainstream research on ice sheet beds. The discovery of MSGL beneath the Rutford Ice Stream by King et al. (2009) was presented as further evidence to discredit the megaflood theory. A second theory of MSGL formation is the groove-ploughing hypothesis (Clark et al., 2003). This theory proposes that MSGLs are grooves in a soft sediment layer (Clark et al., 2003). Large bumps are created in the ice sheet as it passes over uneven hard bedrock upstream, and these "keels" maintain their shape and plough through soft sediments down-ice to produce MSGLs (Clark et al., 2003). Initial modelling suggested that ice keels could survive downstream propagation, (Clark et al., 2003). This evidence has been used to support the groove-ploughing theory (Clark et al., 2003), but it is clear that this hypothesis requires further investigation.

3.2.2.4 Boothia-type erratic dispersal trains

Down-ice transport of a belt of glacial debris from a source area is known as a dispersal train (Dyke and Morris, 1988). In their 1988 study Dyke and Morris proposed that two types of dispersal train exist: a Boothia-type and a Dubawnt-type. Because the Dubawnt-type can be formed by slow ice sheet flow, only Boothia-type dispersal trains are important here. In Boothia-type dispersal trains, debris is dispersed down-ice from a small part of a relatively large source area, and this greater transport of material within the train is proposed by Dyke and Morris (1988) to indicate a zone of ice streaming.

3.2.2.5 Abrupt lateral margins

Ice streams are bordered by slower-moving ice (Stokes and Clark, 2001) and hence former ice streams can be expected to have a sharp zonation at their lateral margins (Stokes and Clark, 1999; Everest et al., 2005). Within an active ice stream setting, Echelmeyer et al. (1994) demonstrated that an abrupt change in velocity at an ice stream margin might create a zone of heavy crevassing. Palaeo-ice streams can be expected to exhibit a similarly proportioned abrupt marginal area (Stokes and Clark, 1999). Examples of identified palaeo-ice streams with abrupt lateral margins include the M'Clintock Channel Palaeo-Ice Stream (Clark and Stokes, 2001) and Ungava Bay Palaeo-Ice Streams (Jansson et al., 2003) in Canada, and the Tweed (Everest et al., 2005) and Strathmore Palaeo-Ice Streams (Golledge and Stoker, 2006).

3.2.2.6 Lateral shear margins

Subglacial ridges, known as lateral shear moraines, have been identified at the margins of palaeo-ice streams (Dyke and Morris, 1988; Hodgson, 1994; Stokes and Clark, 2002a). These features are thought to mark a shear zone separating fast ice flow from the surrounding ice sheet (Dyke and Morris, 1988). The formation mechanism responsible for these features is unknown (Hindmarsh and Stokes, 2008). A number of possible mechanisms for their formation have been proposed (Stokes and Clark, 2002a). These included: meltwater processes depositing sediment in englacial and subglacial streams; melt-out and deposition of entrained englacial debris; downstream sediment recycling; differential erosion; and lateral advection of sediment towards the margin (Stokes and Clark, 2002a). Lateral shear moraines formed part of the evidence for identifying a number of palaeo-ice streams, including the M'Clintock Channel Ice Stream (Clark and Stokes, 2001); and the Strathmore Ice Stream (Golledge and Stoker, 2006).

3.2.2.7 Evidence of pervasively deformed till

The importance of deformable sediment for fast flow has been discussed above. Multiple authors have suggested that ice stream position may well be dependent on subglacial geology and in particular the presence of a soft sedimentary basin (Anandakrishnan et al., 1998; Studinger et al., 2001), thus implying that areas of pervasively deformed till may predispose a section of an ice sheet to fast ice flow (Stokes and Clark, 1999). The presence of deformed till was found in association with MSGL on the bed of the Marguerite Trough Palaeo-Ice Stream (Ó Cofaigh et al., 2005) and in troughs situated in the Ross Sea that were occupied by ice streams (Mosola and Anderson, 2006), and also on palaeo-ice stream beds in the southwest LIS (Evans et al., 2008). Most palaeo-ice streams found so far have been based on evidence of subglacial geomorphology (e.g. MSGLs; Clark, 1993, 1994; Stokes and Clark, 1999, 2001, 2002b) or via associations with large topographic troughs and trough mouth fans indicative of rapid and voluminous sediment delivery (Ó Cofaigh et al., 2003). The signature of fast glacier flow or ice streaming in terms of terrestrial ice-marginal suites of sediments and landforms is much less well known and based largely on interpretations of a small number of Quaternary sedimentary sequences.

3.2.2.8 Submarine till deltas or sediment fans

Large sediment depo-centers on a continental slope may also be indicative of ice stream activity (Stokes and Clark, 1999). Known as trough mouth fans, they are described as fan-shaped, diamict-dominated sediment accumulations (Ó Cofaigh et al., 2003) likely formed as large volumes of sediment are delivered to the shelf edge by marine-terminating ice streams (Ó Cofaigh et al., 2003). Trough mouth fans have been identified in the Polar North Atlantic and on the Antarctic continental margin (Kuvaas and Kristoffersen, 1991; Vorren and Laberg, 1997; Ó Cofaigh et al., 2003). Furthermore, Evans et al. (2012) referred to arcuate morainic assemblages on the prairies as terrestrial equivalents of trough mouth fans (see Section 3.2.2.9), thus these features may not be isolated to marine terminating ice streams.

3.2.2.9 Supplementary evidence for palaeo-ice streaming

There are several features associated with palaeo-ice stream activity that are not identified within Stokes and Clark's (1999) criteria and are perhaps unique to terrestrially terminating ice streams (see Section 3.2.2). One example of this is extensive hummocky topography, which has been reported from areas of former ice streaming at the southern margin of the LIS (Patterson, 1997; Evans et al., 2008). The outer 30 km of many of the margins of the Des Moines Lobe are composed of such terrain, suggesting this lobe underwent periodic stagnation in its outer margins (Patterson, 1997). Furthermore, thrust block moraines have also been reported from the southern margin of the LIS (Evans et al., 2008), especially in southern Alberta, where Evans et al. (2008) suggested that they were large thrust block moraines were formed by the retreating lobate margin of palaeo-ice streams.

Additionally in the absence of a marine environment in which ice stream margins calve and deposit substantial glaciomarine depocentres like trough mouth fans (Vorren and Laberg, 1997), multiple authors have proposed that terrestrial ice streams are likely to fan out to produce lobate snouts (Clark and Stokes, 2003), where subglacial sediments are deposited as tills in down-ice thickening wedges (which form where balance velocity decreases down-flow and debris inputs exceed outputs) (Boulton 1996a, b). Glacial marginal till thickening and stacking of till wedges have been reported from the southern margin of the LIS. Evans et al. (2008, 2012) report large arcuate assemblages of moraines and thick complexes of tills and associated glaciogenic sediments in southern Alberta that demarcate the lobate termini of palaeo-ice streams. This supports Boulton's (1996a, b) theoretical models of till depositional patterns (Fig 3.4). Boulton's (1996a, b) model proposes that the large scale architecture of tills is a direct result of the fact that the tills deform and are strongly coupled with the glacier. This coupling ensures that erosion or net loss of subglacial sediment occurs in the accumulation area where ice is undergoing acceleration. In contrast, in the ablation zone where ice is decelerating, deposition of deforming layer sediments takes place. This produces a pattern of till deposition whereby till thickens towards the sub-marginal zone of the glacier. Although more recently this model has been questioned based upon till architecture in Germany (Piotrowski et al., 2001), stacked sequences of tills do occur at modern glacier snouts where they constitute push/squeeze moraines constructed during periods of ice standstill (Krüger 1993, 1994; Evans and Twigg, 2002). This shows some support for Boulton's (1996a, b) model. These temperate glacier snouts have high erosional capacities, and effectively transport large volumes of subglacial sediment from an erosional zone to a depositional zone a few hundred meters wide (Boulton, 1987).

Alley et al. (1997) also suggest that the wide range of sediment transport mechanisms (net adfreezing, supercooling, folding, thrusting and the concentration of subglacial fluvial sediment) in a glacier become increasingly active in marginal and sub-marginal locations, resulting in the concentration of debris. Additionally, Alley et al. (1997) point out that continuous deforming beds are most likely to occur in ice-marginal locations, especially if pre-existing sediment (glaciolacustrine/glaciomarine sediments) blankets bedrock obstacles. In contrast to the extensive literature on the deposits associated with marine based ice stream margins, there are few case studies of the associated sediment-landform assemblages of terrestrial-terminating ice streams. Thus further investigation is needed to test the theoretical models established by Boulton (1996a, b) before such features can be proposed as diagnostic of land terminating ice streams.

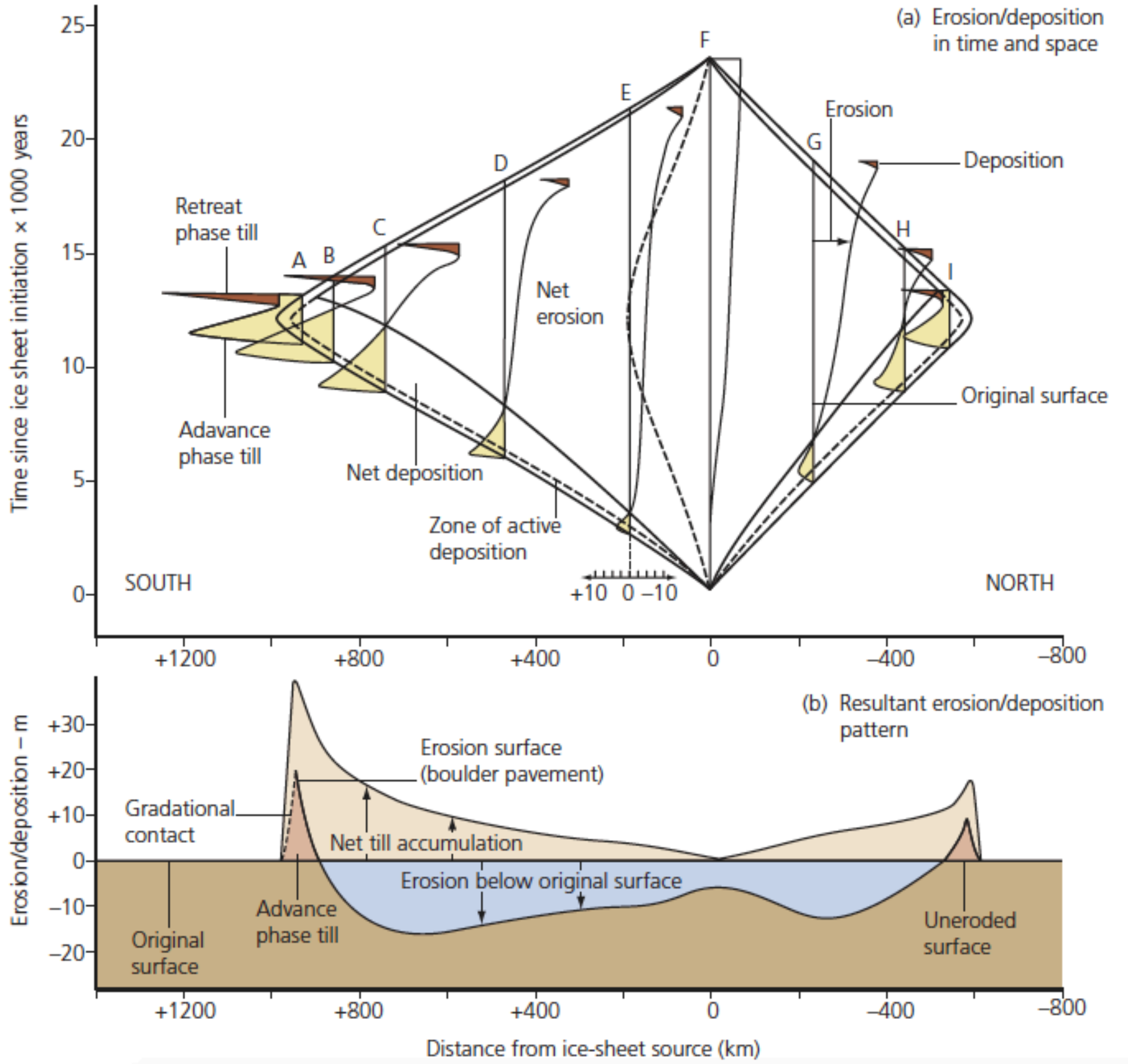


Figure 3.4: ***A.*** *Time-distance diagram showing conceptually modelled shifts in the patterns of subglacial erosion and deposition associated with the evolution of an ice sheet. Graph line A-1 represent deposition and erosion at selected locations.* ***B.*** *Resultant pattern of deposition and erosion produced by a single glacial cycle. Originally produced by Boulton (1996a) and reproduced by Benn and Evans (2004).*

3.3 Summary

Ice streams are fast-flowing corridors of ice that have a key role to play in ice sheet stability (Stokes and Clark, 1999, 2001; Bentley, 2003). The relative inaccessibility of contemporary ice stream beds means that locating palaeo-ice stream tracks is of great importance for improving understanding of the processes and dynamics in operation (Stokes and Clark, 1999, 2001; Ó Cofaigh et al., 2005). Multiple studies have outlined geomorphological evidence associated with palaeo-ice streams, from which Stokes and Clark (1999) proposed eight diagnostic criteria. However very few studies review the sediment and landforms associated with terrestrially terminating ice streams, highlighting the importance of further investigation of this type of palaeo-ice stream.

4. Methods

This chapter reviews the methods used to address the research aims and is subdivided into two sections, consistent with the two approaches used to investigate the characteristics and distribution of glacial sediments and landforms. Section one reviews the methods employed in mapping glacial geomorphology in the SWSS. The second section describes the stratigraphic analyses used and its benefits in understanding ice stream dynamics. This section also documents the application of Rockworks 16™ in three-dimensional subsurface stratigraphic modelling. A supplementary user guide detailing the step by step process of modelling stratigraphic units in Rockworks 16™ is then reproduced in Appendix 1.

4.1 Geomorphological mapping

In order to investigate the geomorphology of the SWSS and to build on the preliminary mapping previously undertaken by Ó Cofaigh et al. (2010) and Ross et al. (2009), mapping in this study was undertaken at a higher resolution and at the larger scale of 1:10,000, using a combination of space borne imagery (SRTM, and Landsat ETM+). This facilitated not only accurate mapping of landform components but also an assessment of their inter-relationships as a palaeo-ice stream landsystem.

4.1.1 Shuttle Radar Topography Mission (SRTM) 3-arc second (90 m) DEM

3 arc second (90-m resolution; Jarvis et al., 2008) Shuttle Radar Topography Mission (SRTM) imagery was employed to create DEMs of the landform record. While other available topographic data sets exist (ASTER, CDED), SRTM3 was chosen as it was considered the best available data set for geomorphological mapping. ASTER (Advanced Spaceborne Thermal Elevation and Reflection Radiometer) provides data with higher or similar resolution (15-90 m) to SRTM however, the data is only available on demand and requires payment. Due to the financial constraints of this research, ASTER data has been dismissed as a mapping technique. Canadian Digital Elevation Data (CDED) provides topographic information with a resolution varying between 0.75-3 arc seconds and is referenced to the National Topographic Data Base (NTDB). When CDED data was examined contour data was easily visible throughout the image, so hindering the mapping and interpretation process. Thus this topographic data set was also dismissed.

The SRTM was flown in February 2000, obtaining elevation data on a near global scale (60°N - 56°S) (Ramirez, 2006). The mission produced three data sets: 1 arc-second, 30 m resolution data set for the US only (SRTM1); 3 arc-seconds, 90 m resolution (SRTM3) and 30 arc-seconds, 1 km resolution (SRTM30) data set with near global coverage (NASA, 2001). All three data sets have an approximate swath width of 225 km and the data is provided in an '.hgt' file format, which is a 16-bit signed integer data in a simple binary raster (NASA, 2001). All elevation data is in metres and referenced to the World Geodetic System (WGS84).

4.1.1.1 SRTM manipulation in Global Mapper™

It was necessary to convert the SRTM file format '.hgt', as it was only recognised by a limited number of GIS packages. Global Mapper™ was used to produce a smoothed render image of the SRTM data that could be manipulated to accentuate features, produce 3D images and change illumination angles. By vertically exaggerating the elevation data, it was possible to more easily identify landforms as they become more defined. However it must be noted that because vertical exaggeration stretches the data above a set height (the lowest point in the data set= 460 m) this method can start to misrepresent features due to morphology changes. Thus a maximum vertical exaggeration of 0.025% (Fig 4.1) was used and images produced in this way were not used as the sole means of mapping but rather to help interpret and distinguish landforms.

As proposed by Smith and Clark (2005) to minimise azimuth bias, multiple illumination angles were also used to aid mapping. With respect to the study area, where there is a myriad of landforms all of varying orientation and scale, it was necessary to implement illumination angles orthogonal to each other and when mapping flow-sets parallel and normal to the main lineation direction (Smith and Clark, 2005). Finally, 3D images were created (Fig 4.1) to help verify mapping data, as this enabled multiple viewing angles of the same images.

4.1.2 Landsat ETM+

The Landsat 7 satellite consists of 8 spectral bands, which record earth surface radiation. Bands 1-5 and 7 have a 30 m resolution and allow colour composites to be created to aid image interpretation. Band 6 (thermal infrared) has a resolution of 60 m. The satellite is also fitted with an Enhanced Thematic Mapper Plus (ETM+), which provides an extra panchromatic band (band 8). This band produces images with a spatial resolution of 15 m (Lillesand et al., 2008). Each scene size is ~170 km north-south by 183 km east-west (USGS, 2006).

Additional geomorphological mapping was conducted through interpretation of the panchromatic band (band 8: 0.52-0.90 μm) due to its high resolution (Fig 4.2). A total of 8 scenes were mosaiced (Table 1). These were downloaded from the Global Land Cover Facility (GLFC) website (hhtp://glcf.umd.edu) and were cropped to the size of the study area within ArcMap 10.3. All images were in GeoTIFF format and have been georeferenced with the North American Datum of 1983 (NAD83) corresponding to the Universal Transverse Mercator (UTM) projection; Zone 12.

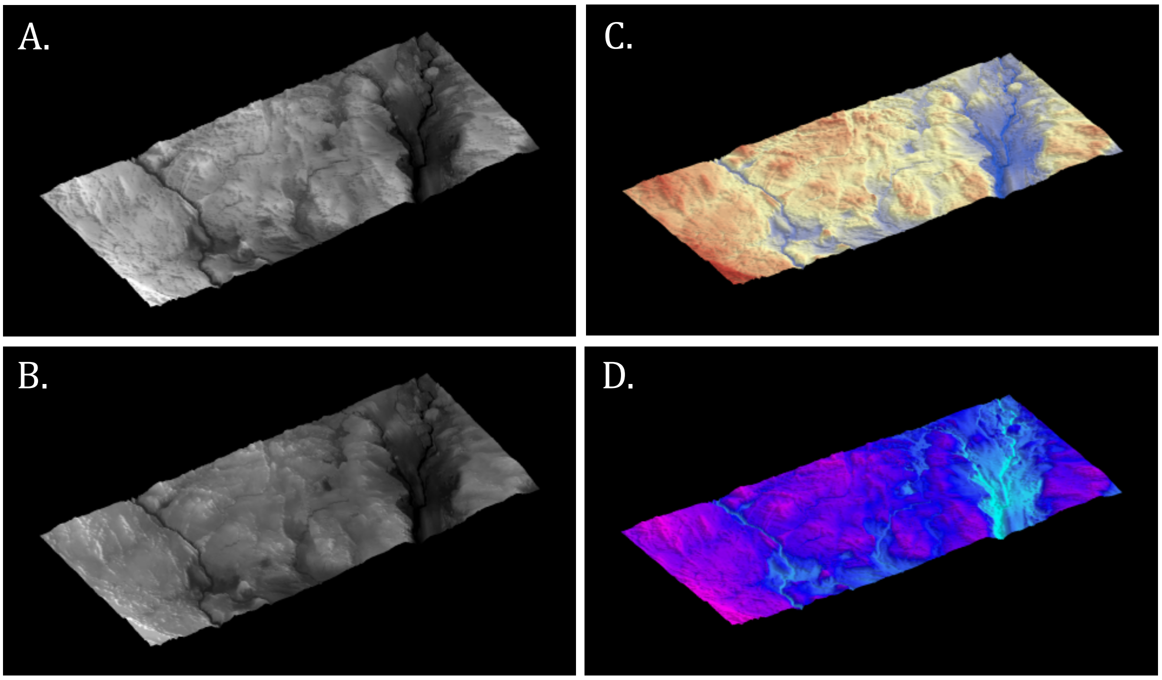


Figure 4.1: *3D images created in Global Mapper™ of the study area. The 360° enabled multiple views of landforms, allowing the differences in relief to be more easily identified. Images are vertically exaggerated by 0.025% to further aid landform identification.* ***A/B.*** *provide two illumination perspectives. Note the distinct visual changes in the imagery from a single DEM manipulation.* ***C/D****. Provide two different pseudo-colour images. Changes in pseudo-colors proved to be of limited use when identifying subtle landforms. Rendition is for illustration purposes only and no scale is given.*

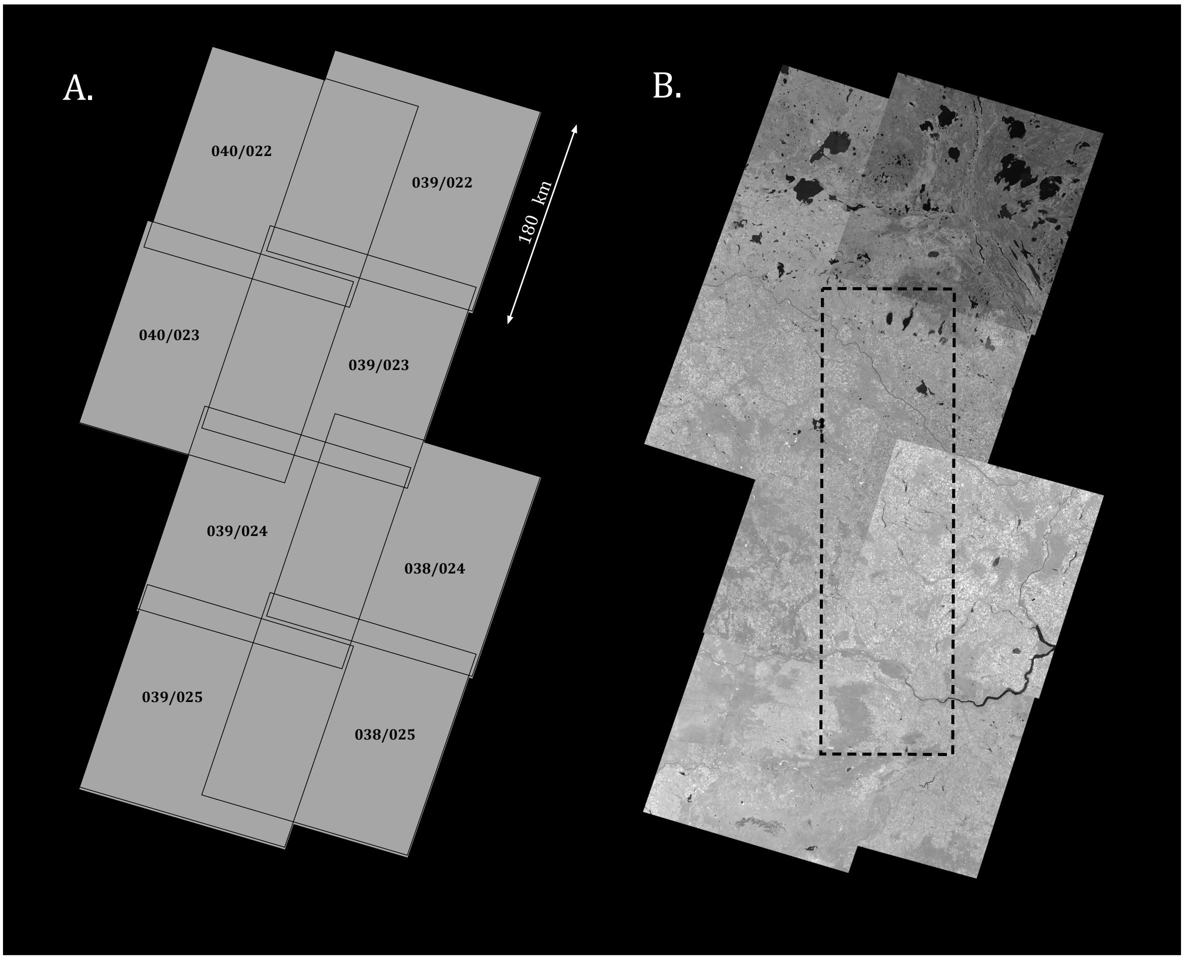


Figure 4.2: *Landsat ETM+ imagery of the study area.* ***A.*** *The eight individual Landsat ETM+ tiles covering the study area.* ***B.*** *Mosaic of Landsat ETM+ tiles (Band 8). Dashed black box corresponds to study area. Rendition is for illustration purposes only and no scale is given.*

Table 4.1: *Details of the eight Landsat tiles used, including sensor type, path, row and acquisition date.*

Sensor type	Path	Row	Acquisition date
ETM+	*038*	*024*	*29/07/1999*
ETM+	*038*	*025*	*03/08/2001*
ETM+	*039*	*022*	*22/09/1999*
ETM+	*039*	*023*	*23/09/2000*
ETM+	*039*	*024*	*23/09/2000*
ETM+	*039*	*025*	*22/07/2000*
ETM+	*040*	*022*	*17/08/2001*
ETM+	*040*	*023*	*10/10/2001*

4.1.3 Mapping techniques compilation

Mapping from SRTM imagery (Fig 4.3) was conducted by on-screen digitisation of features within ArcMap 10.3; ArcMap 10.3 was used to digitise landforms in preference to Global Mapper™ as it is difficult with the latter to easily digitise complex glacial geomorphology. All geomorphology was mapped at a variety of scales in order to avoid bias within the mapping process. Individual landforms where identified, within the imagery, by a particular pattern of reflectance. For example, glacial lineations where identified typically in groups with contrasting reflectance on each side of their linear crest line. Whereas, major moraines where identified based on their relatively strong crestline and in many cases their reflectance differed strongly on each side (with a more gradual change in reflectance on one side). Features were mapped as separate shapefiles either as polygons or lines. Different features were digitised on separate vector layers after visual interpretation (e.g. Clark, 1997) and stored as shape files (.shp) (Fig 4.4). The same method was adopted for mapping the Landsat ETM+ images (Fig 4.3) as was used to map from the SRTM imagery. The images were opened in ArcMap and overlaid with the same vector layers that were used to map the DEMs. This allowed first order verification of the SRTM interpretations. Whilst the Landsat ETM+ and SRTM data sets are georeferenced to different co-ordinate systems no problems were encountered when overlaying vector files on the different raster formats. The colour composite Landsat images created by alternating band combinations provided no advantage over the higher resolution panchromatic band. This is most likely due to the 30 m resolution and the extensive agriculture and subsequent modification of subtle landforms, requiring higher resolution imagery to identify such features. The 15 m resolution of the panchromatic band enabled more detailed mapping of smaller features but the lack of any topographic information hindered the identification of the larger, regional scale landforms. Band 8 was significantly more useful when identifying smaller lineations and subtle differences within sections of hummocky terrain than SRTM.

Using a combination of SRTM and Landsat EMT+ imagery, landforms were and mapped at a variety of scales to avoid any scale bias in their detection. Upon completion of mapping, vector layers were complied (.esp files) into Adobe Illustrator CS4. The final map (see Map Sheet 1 and 2) is intended to be viewed at A1 size.

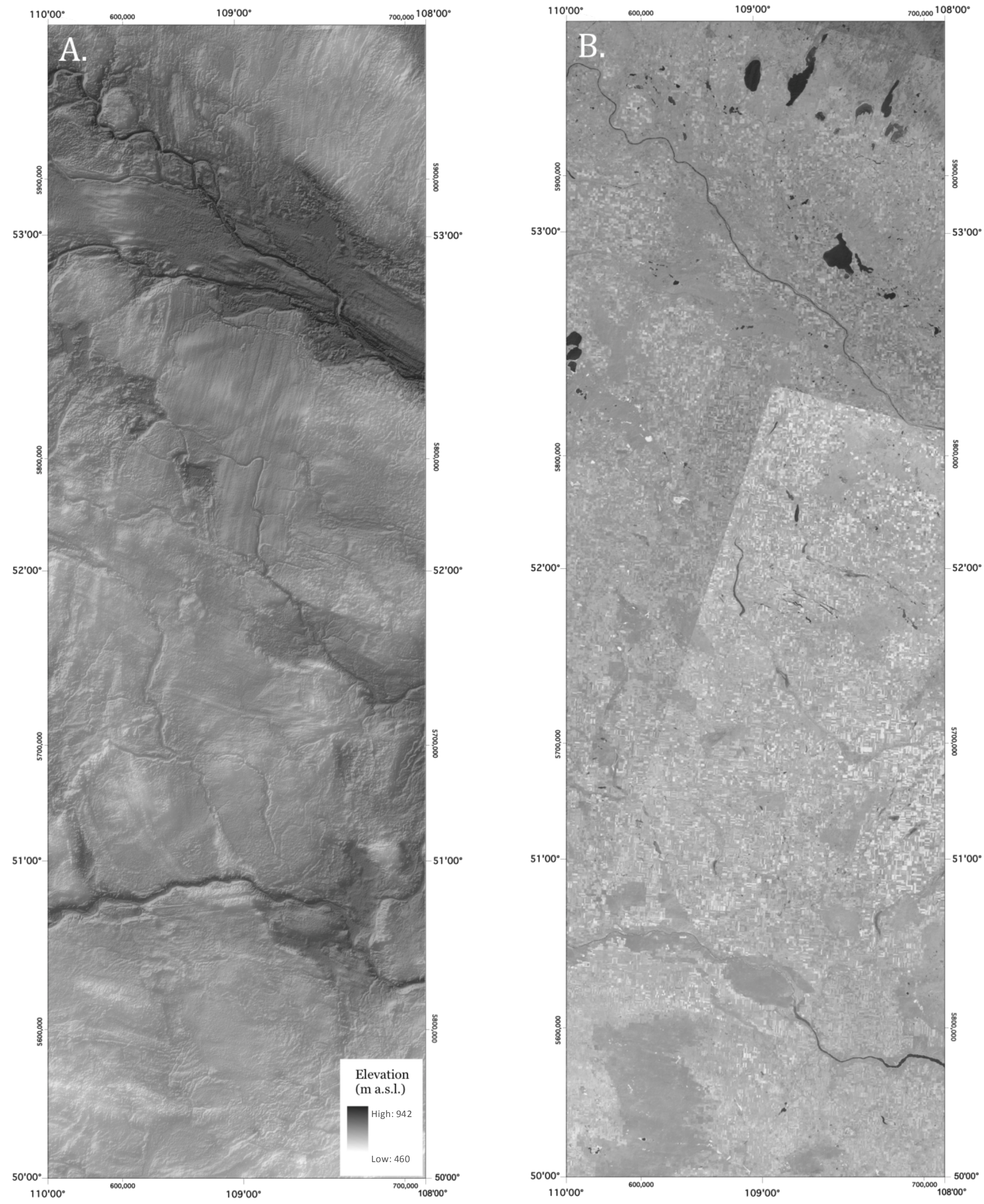


Figure 4.3: *Exported Geotiffs of the study area used for geomorphological mapping.* ***A.*** *SRTM-90 DEM (note several Geotiffs of this image with variable illumination angles and multiple pseudo colour combinations were also used). Black represents the lowest ground and white represents the highest ground* ***B.*** *Landsat ETM+ imagery of study area.*

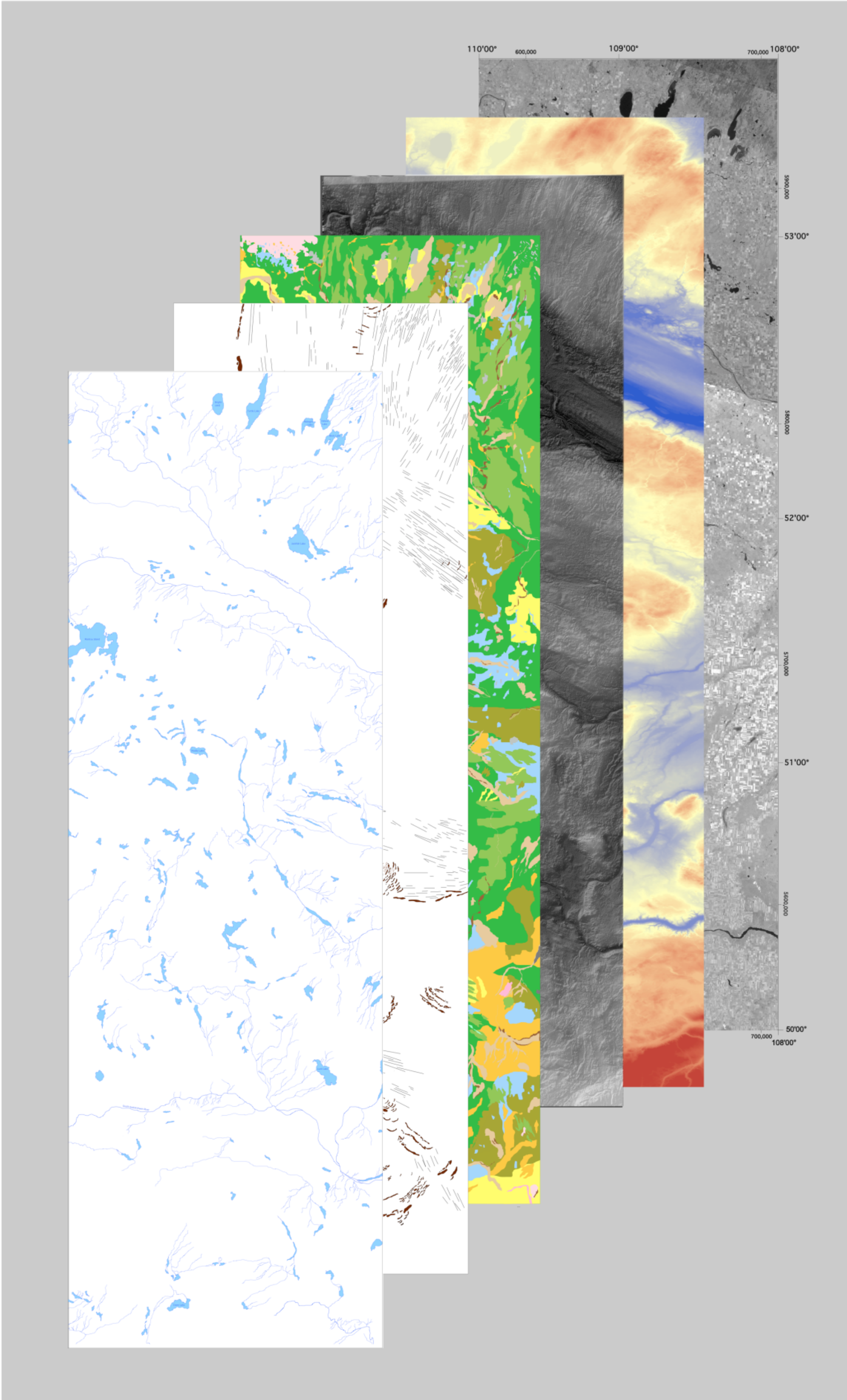


Figure 4.4: *Simplified summary of layered shapefiles and imagery used for geomorphic mapping. From left to right: Sample polygon and line shapefiles, surficial geology, SRTM imagery as Geotiffs (various illumination angles and pseudo colour schemes, Landsat ETM+ imagery.*

4.2 Stratigraphic analysis

The following section describes the use of subsurface stratigraphic data in this study (Fig 4.5). As mentioned above Rockworks 16™ three-dimensional geological modelling software was used to interpret continuous stratigraphic surfaces, which were stacked to form a 3D stratigraphic model. Other functions enabled the creation of individual logs and multiple log sections, fence diagrams and isopach maps (Rockworks Inc, 2013). This specialised GIS modelling software has been used relatively little in the context of glacial sedimentology thus the program's overall system architecture is described below. For details of the exact tool, processes and modelling techniques used refer to Appendix 1.

4.2.1 Data compilation

Subsurface geological data was compiled from boreholes (auger holes) drilled by government agencies (Saskatchewan Research Council, Saskatchewan Department of Agriculture and Saskatchewan Institute of Pedology), and multiple private oil companies (Fig 4.6). 197 continuously cored boreholes to depths >288 m below ground surface were used to create a subsurface database covering the 57,400 km^2 study area. Boreholes were drilled between 1962-1978 and information obtained consists of electric logs (drilled through Quaternary deposits into bedrock), field log descriptions, and information from drill cuttings and sidewall samples.

Borehole locations were established from borehole records, the distribution of which is shown in Figure 4.6. Coordinates were converted from the Alberta Township System (ATS) to Universal Transverse Mercator (UTM) locations in (x) and (y). Borehole records, containing coded information and driller descriptions regarding sediment characteristics (see Section 4.2.3), the elevation of changes, unique ID's and a start (collar) height (z), were entered into an Excel spreadsheet. Records were then reformatted for entry into Rockworks to form a digital database composed of two file types; index files and log files (Table 4.2) (Rockworks Inc, 2013). Using the DIGIDATA extension, logging curves (self-potential and electrical resistance) were digitised and then recreated using the LOGPLOT extension to aid in lithologic interpretation (Rockworks, Inc, 2013) (see Fig 4.7 for example).

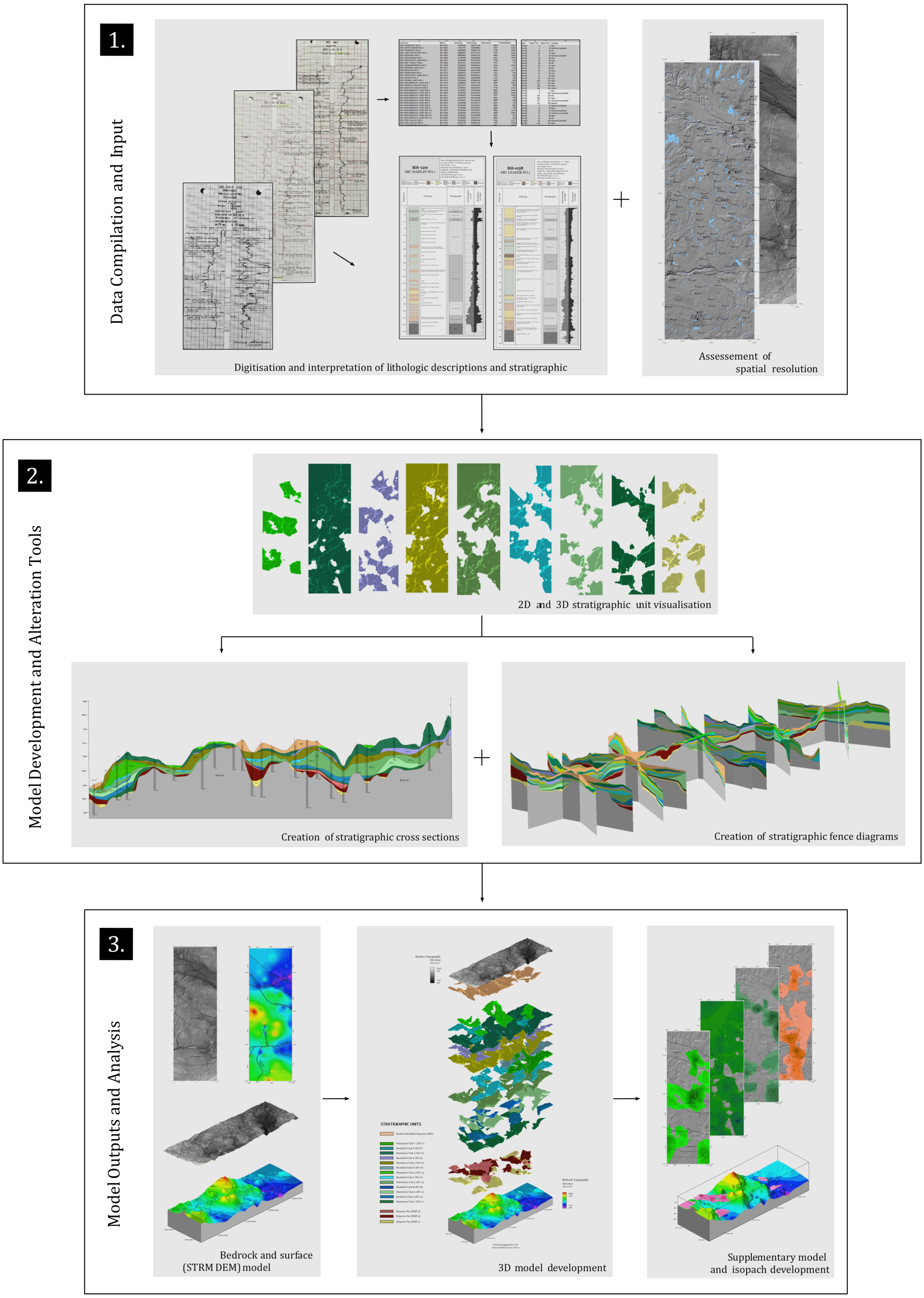
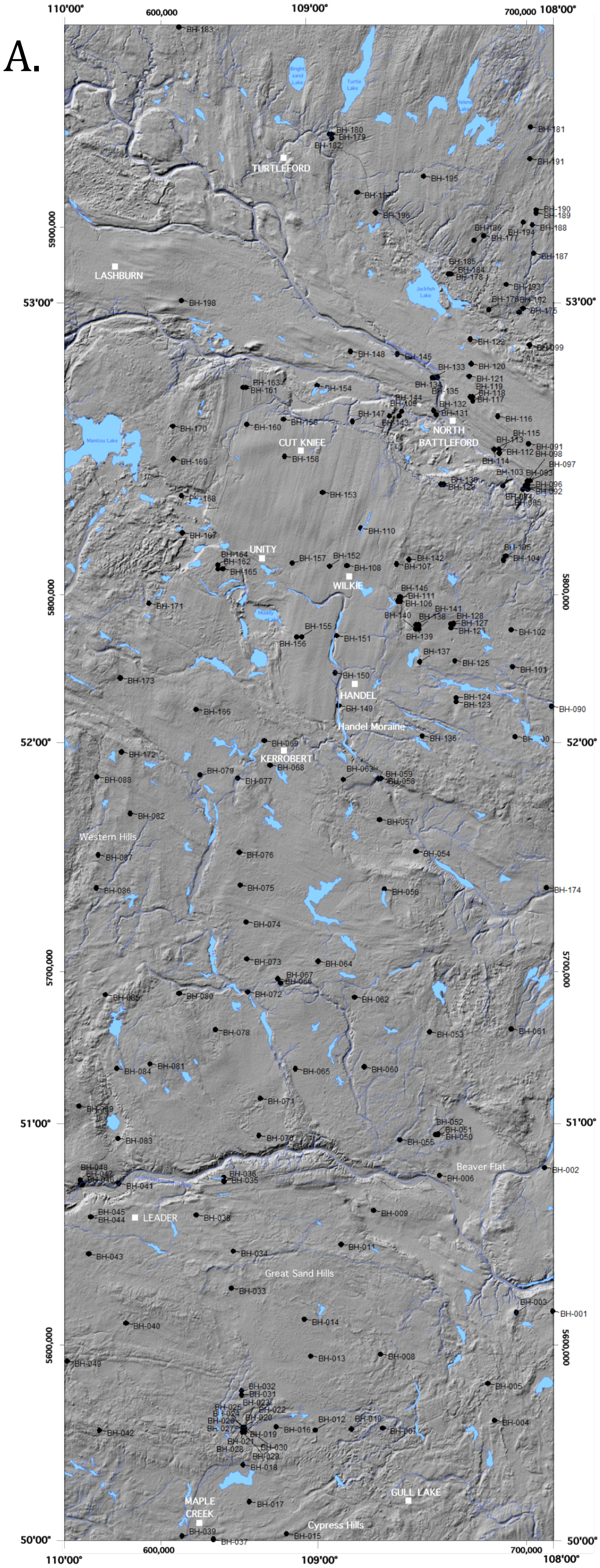


Figure 4.5: *Summary of sedimentary analysis system architecture used in this study. Each section corresponds to text descriptions below. See Appendix 1 for detail of imagery and model production.*

Figure 4.6: ***A.*** *Distribution map of borehole locations within the SWSS. Each point relates to a single borehole location at which electric logs (drilled through Quaternary deposits into bedrock), field log descriptions, and information from drill cuttings and sidewall samples were obtained. Boreholes range from BH-001 to BH-198.* ***B.*** *Regions referred to in reference to borehole locations.*



B.

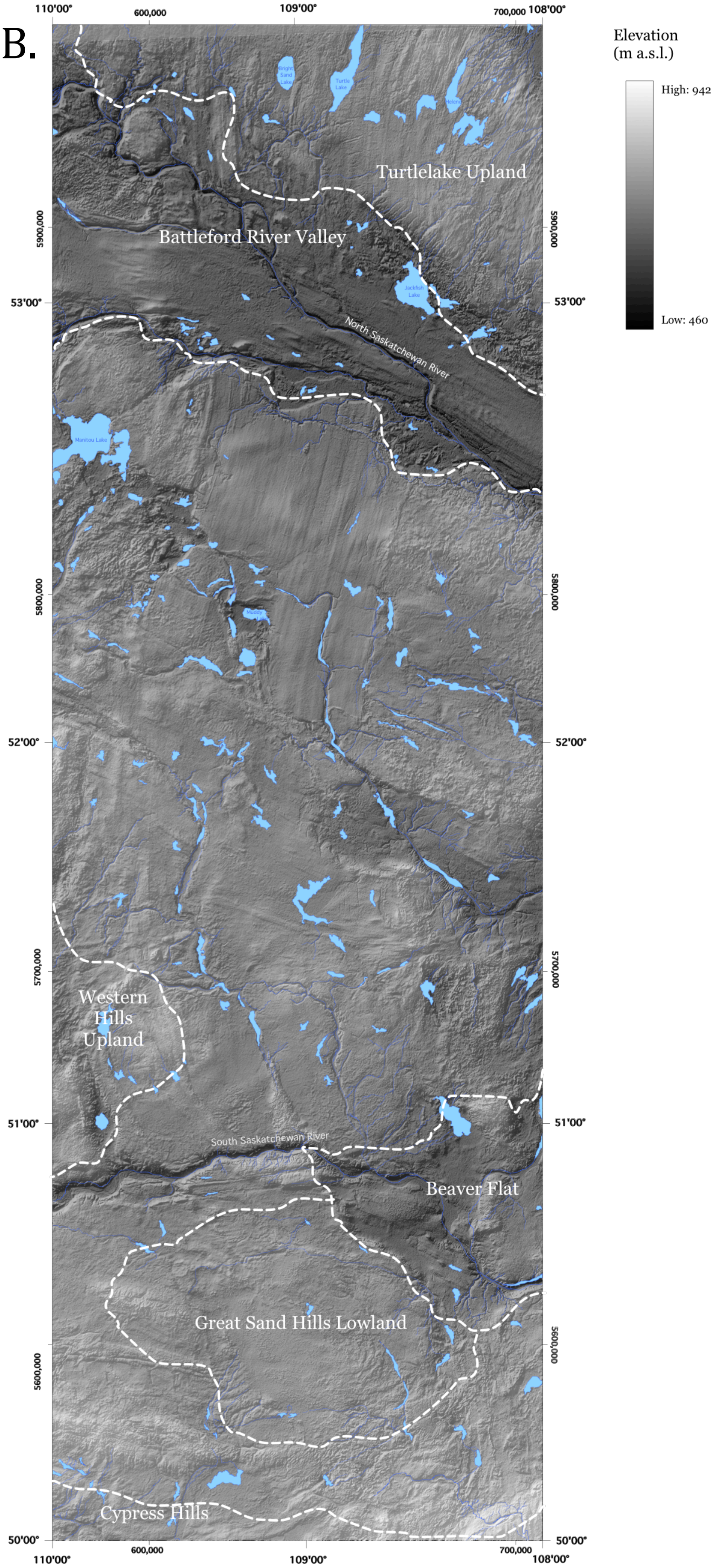

Elevation
(m a.s.l.)
High: 942
Low: 460
Turtlelake Upland
Battleford River Valley
North Saskatchewan River
Western
Hills
Upland
South Saskatchewan River
Beaver Flat
Great Sand Hills Lowland
Cypress Hills
110'00°
109'00°
108'00°
600,000
700,000
53'00°
52'00°
51'00°
50'00°
5900,000
5800,000
5700,000
5600,000

Table 4.2: *Examples of reformatted index and log files used to import data to Rockworks 16™*

Unique Borehole ID	Original borehole name	Easting (x)	Northing (y)	Start (collar) height (z)	Total depth
BH-001	*SRC PENNANT NO.1*	*704599*	*5607110*	*680*	*141*
BH-002	*SRC WHITE BEAR NO.1*	*702375*	*5645993*	*610*	*130*
BH-003	*SRC PENNANT NO.2*	*694689*	*5606733*	*718*	*189*
BH-004	*SRC LAKE ANTELOPE NO.1*	*688642*	*5577571*	*724*	*143*
BH-005	*SRC ROSERAY NO.1*	*686864*	*5587525*	*684*	*148*

Unique Borehole ID	Depth start of unit (meters)	Depth to base of unit (meters)	Lithology
BH-001	*0*	*5*	*Clay*
BH-001	*5*	*7*	*Silt clay*
BH-001	*7*	*32*	*Diamicton (oxidised)*
BH-001	*32*	*47*	*Diamicton (unoxidised)*

4.2.2 Interpretation of drillers' lithologic descriptions

The methods used to interpret drillers' logs are a modification of lithofacies analyses used in sedimentary studies (Miall, 2015). The drillers' descriptive logs were imported into Rockworks by translating the inconsistent and informal descriptors used to a consistent set of lithologies. For example, the clay category included tough clay, brown clay and clayey deposit. Compound descriptions were broken down into fractions for tabulation. Silt and clay, for example, were counted as 50% of each for the described interval. Silty clay was assumed to be 25% silt and 75% clay for the described interval. Following this pattern, clayey silt was assumed to be 25% clay and 75% silt. The resulting consistent set of lithologies allowed for the conversion of hundreds of unique descriptive terms, to a limited vocabulary of ~35 lithologic keywords, which were considerably more suitable for use in geological modelling.

4.2.3 Identification and differentiation of stratigraphic units

Rockworks was used to create unique composite lithologic striplogs for all 197 boreholes, which could then be compared and interpreted. Stratigraphic assignments for the materials in the striplogs were recorded in the Rockworks spreadsheet.

Stratigraphic units consist of one or more lithologic units that are closely associated in vertical succession, have similar textural characteristics, and were probably deposited under similar environmental conditions (Nichols, 2009). While a stratigraphic nomenclature does exist for Saskatchewan (see Section 2.3.2), rather than trying to fit observations into the existing stratigraphic framework, which could easily result in over interpretation of the sequence stratigraphy, stratigraphic units were designed informally as till units 1-7, intertill/stratified units A-F and Empress Group units 1-3. This facilitated accurate descriptions and interpretations of the Quaternary succession in terms of lithologic properties, thickness, distribution, and genesis. Within the discussion section (see Section 7.1.4) an attempt is then made to correlate these units with those previously defined in the region (Whitaker and Christiansen 1972; Christiansen 1992; Barendregt et al., 1998).

The differentiation and identification of stratigraphic units within this study was accomplished via careful correlation of: **1.** lithological properties including, colour, grain size of the smaller than-gravel fraction, structure, stratification, partings, inclusions, and the nature of contacts; **2.** geophysical properties, recorded by electrical resistivity and self potential. Detailed description of analyses of these properties and a description of the application of these properties in defining units containing glacial diamicton are provided below. Figure 4.7 is an example of such plots of the data from BH-038 in which units containing glacial diamicton are recognised. In most places, the contact between the units are easily defined by abrupt changes in the majority of properties, and in some places the units are separated by stratified (non glacial) units. However in a few areas the contacts are more gradual and only an approximate boundary can be defined. In such cases the boundary was defined as the depth at which there is the greatest change in composition.

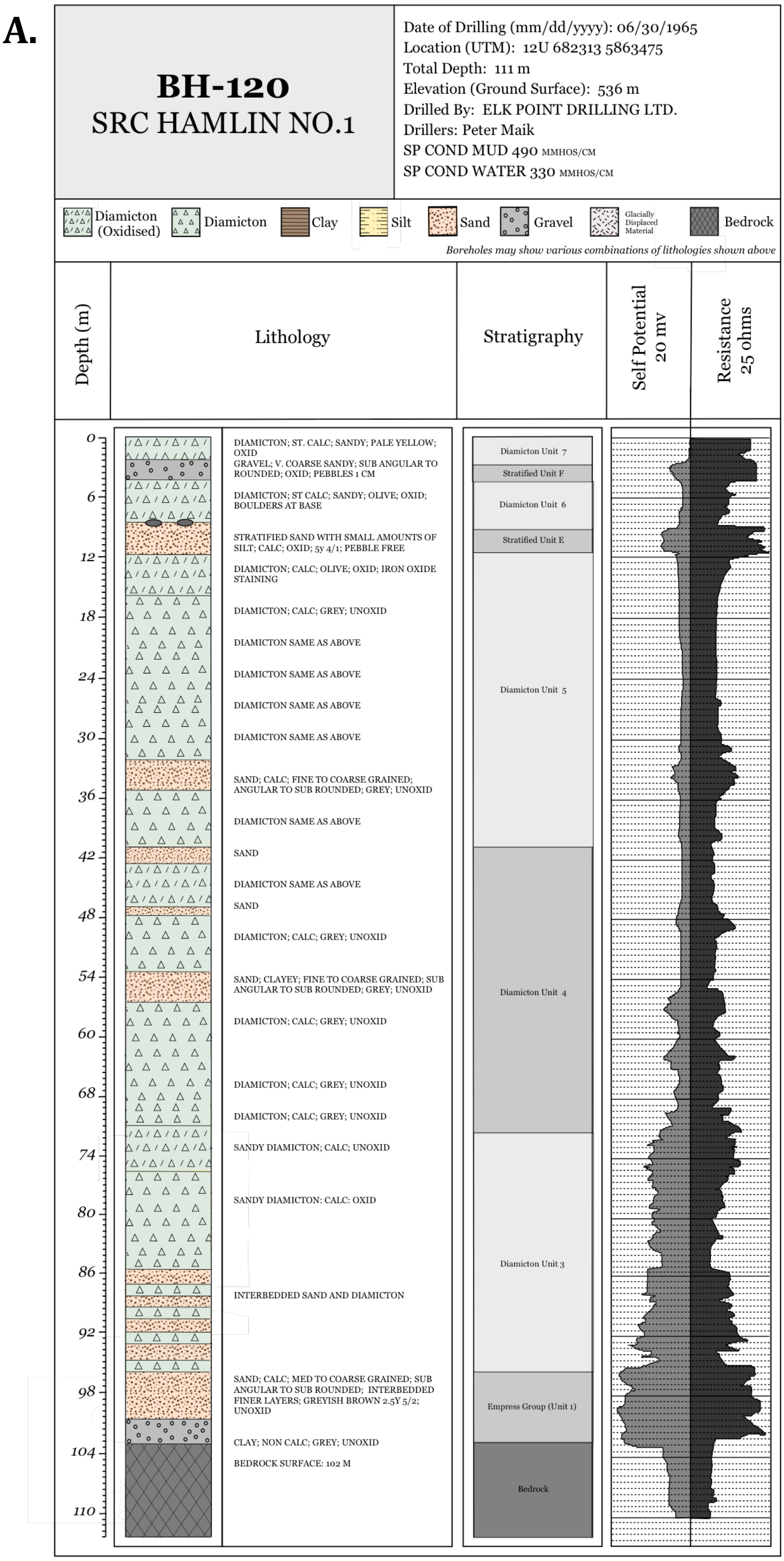
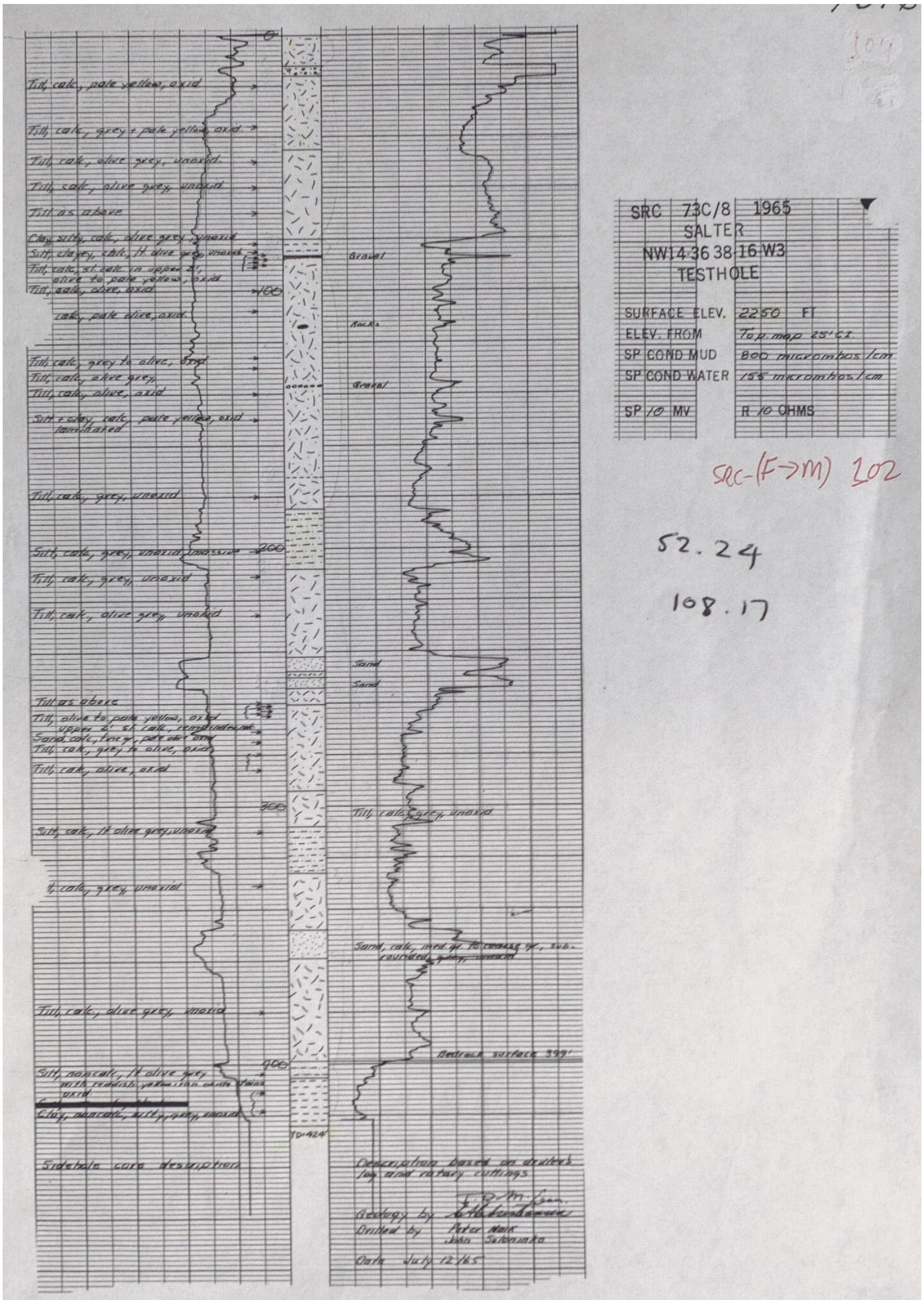


Figure 4.7: *Example of a composite borehole log from the SWSS.* ***A.*** *Log is created using both DIGIDATA and LOGPLOT extension in Rockworks 16™ and has been manipulated and finalised in Adobe Illustrator CS4.* ***B.*** *Original drillers' descriptive log.*

B.



SRC 73C/8 1965
SALTER
NW14-36 38-16-W3
TESTHOLE
SURFACE ELEV. 2250 FT
ELEV. FROM Top map 25' CI.
SP COND MUD 800 micrombos/cm
SP COND WATER 155 micrombos/cm
SP 10 MV
R 10 OHMS
SRC-(F->M) 102
52.24
108.17
Sidehole core description
Description based on electric log and rotary cuttings
Bedrock surface 399'
Date July 12/65

4.2.3.1 Lithologic properties

Colour- Colour was recorded (in a moist state) for both weathered and oxidised, and unweathered and unoxidised sediments. Colours were recorded on borehole logs using Munsell colour charts and coding system. Weathered units in the study area were commonly very light brown, compared to the dark grey of unweathered units. This is probably a result of oxidation of iron-oxide particles in the sediment (Andriashek and Fenton, 1989). Units that showed oxidation through the entire sample and not only along fracture surfaces were referred to in this study as being oxidised. In most places these units were interpreted to have been exposed to surface weathering after deposition, especially where unoxidised till overlies oxidised till; in such places the oxidised zone is considered to represent a weathered surface on the lower till.

Grain size- Recorded on the majority of drillers' logs are field estimates of the dominant grain size of <2 mm fraction. Related to grain size is the moisture content, which was expressed as dry, moist or wet.

Pebble lithology- On borehole records only easily recognised differences in the pebble lithology were recorded. However in many cases drill sample logs commonly did not contain sufficient pebbles for meaningful counts.

Stratification, partings and laminations- These structural elements were recorded on borehole logs. Bedding and partings were recorded from samples even though they are twisted and distorted by the drilling process.

Inclusions of glacially displaced sediments- Masses of glacially displaced and incorporated older sediment are recorded in some borehole samples where displaced material is lithologically distinct from the host material. In many places these displaced materials are clearly identified on drillers' logs.

Nature of contact- The nature of the contact between units was recorded on the majority of borehole logs. However in a minority of cases these descriptions were very brief and not available for all contacts. In these cases comparisons were made to the surrounding boreholes logs.

4.2.3.2 Geophysical properties

Self-potential and electrical resistance- Electric logs, which record the self potential and electrical resistance of different sediment types, were available at all borehole locations. The electric logs of the stratigraphic units appear to be affected by a number of factors. In general, coarse-grained sediment such as sand and gravel has a high resistance, especially if the sediments are dry. The presence of water and dissolved salts, however, reduces the resistance, thus producing a log spike without a change in lithology. Fine-grained sediment such as silt and clay generally show less of a difference in their signature and have a very low resistance. Lithified silt and clay such as marine claystone and siltstone bedrock types have an even lower electrical resistance and higher self-potential. Presumably this reflects the more consolidated and cemented nature of the rock. For the most part, electric logs recorded in the SWSS were sensitive enough to detect differences in the properties of different tills. Generally, the logs appear to record the major differences in grain size of the till, though other factors such as a high degree of fracturing at the top of a unit or more abundant carbonate within the till also appear to contribute to a higher electrical resistance.

4.2.4 Cross section and map construction

As groups of logs were interpreted and stratigraphic striplogs were created (Fig 4.6), the cross section function on Rockworks was used to produce well-to-well cross sections (Map Sheet 3) (Rockworks Inc, 2013). These cross sections were fundamental to adjusting stratigraphic picks, refining correlations, and improving the understanding of not only the succession of units, but also aerial variations in their thickness and elevation. In order to further check the consistency of the correlations, many ‘closed’ sections were constructed (starting and ending at the same borehole) allowing individual stratigraphic units to be traced back to their original stratigraphic position (Rockworks Inc, 2013).

In addition to cross sectional diagrams, to gain an alternative view of stratigraphic units, 2D surface models of the tops of major stratigraphic units were created to evaluate trends and further check tentative correlations. Computer-generated isopach maps were also reviewed in order to identify anomalous data points. These data points were evaluated and in some cases manually edited in selected places (see Section 7.4.5 for evaluation of the application of Rockworks in creating isopach maps) to remove outliers that were well outside the main part of a unit, resulting in more consistent isopach trends.

4.2.5 3D model development

The final step was the development of 3D geological models. Rather than using the Rockworks software package to create a complete 3D stratigraphic model directly from borehole data via a single model run, individual surfaces were modelled separately then imported into a single 3D grid model (Rockworks Inc, 2013). Thus allowing them to be checked and individual settings to be selected for each surface.

In this type of 3D model production, an inverse modelling algorithm is used to extrapolate numeric codes that represent a single stratigraphic class (Rockworks Inc, 2013). Grid nodes situated surrounding drill holes are assigned a value that corresponds to a stratigraphic class based on the proximity of a grid node to the boreholes that surround it. In this manner, interpretation continues until the program finds a cell that is already assigned a stratigraphic class (Rockworks Inc, 2013). One limitation of this type of numerical interpolation is the sensitivity to the distribution of boreholes, with values from isolated drill holes tending to extrapolate outward to fill an inordinate amount of the model area (Rockworks Inc, 2013). During preliminary modelling this effect was particularly noticeable in the north of the study area. To minimise this problem the study area was cropped from a former 61,000 km^2 area to the 57,400 km^2 to minimise the area in the north where very few boreholes were available.

Grid nodes for 3D interpolation were 500 m horizontally and 10 m vertically in size. This vertical dissection was chosen as it was a compromise between preserving geologic detail, and computational capability such that model runs could be completed in a reasonable time. When all surfaces were completed, the Rockworks 'grid appending' tool was used to build them into a single gridded model. Firstly a blank model of the desired dimensions was created. The lowest surface model (bedrock surface) was then introduced and all nodes with values below this surface were re-coded so that they were equal to the bedrock stratigraphic code. Sequential surfaces up the stratigraphic column were then added to form the 3D model until the land surface was reached. In each case, the corresponding stratigraphic code was used to recategorise blank cells situated beneath the new surface model (Rockworks Inc, 2013). The upper surface of the land surface model was constructed using the core top (collar) elevations (z) of each borehole. Unlike other layers the topography between boreholes was not extrapolated, but rather the SRTM DEM surface was used to constrain this layer's surface. The base of the model was trimmed 25 m below the bedrock surface of the deepest measurement.

4.3 Summary

The methodological approach presented above has been designed to provide a logical and comprehensive assessment of geomorphological and stratigraphic data. This will enable a thorough evaluation of the palaeo-ice stream behaviour within the SWSS. Accurate mapping of glaciological geomorphology, the identification of stratigraphic units, and the construction of a regional stratigraphic model, will collectively allow the depositional patterns associated with ice stream activity to be discussed.

Chapter 5. Descriptions of Geomorphology and Stratigraphy

This chapter is divided into two sections. The first section presents the results obtained from glacial geomorphological mapping of the SRTM and Landsat ETM+ imagery. Section 2, describes the characteristics of surface and subsurface sediments and stratigraphy within the study area.

5.1 Glacial geomorphology

A variety of glacial landforms have been described for southwestern Saskatchewan by Campbell (1986a, b, 1987a, b) and more recently by Ross et al. (2009) and Ó Cofaigh et al. (2010). This section briefly defines the key landforms that have been mapped in this study. The geomorphological map of the study area can be found in the inside back cover of this thesis as Map Sheet 1.

5.1.1 Megageomorphology and lineations

Lineation mapping highlights a prominent corridor of north-south streamlining (Fig. 5.1). This corridor is hereafter referred to as Corridor 1 so that comparison can be made with previous work (Ó Cofaigh et al., 2010). This corridor of smoothed topography and lineations stretching from the northern perimeter of the study area terminates at a dense arcuate series of ridges (TR-1) orientated perpendicular to the lineation long axes. Within its upper reaches the corridor displays a characteristic convergent flow shape where it is ~120 km wide followed by a narrow ~60 km wide trunk zone and a progressively wider terminal zone (170 km) (Fig 5.1). Three smaller corridors of streamlined terrain also occur which cross cut Corridor 1. They run northwest to southeast across the study area. These corridors (hereafter referred to as Corridors 2A, B and C (Fig 5.1)) are narrower, at ~40 km wide, and align perpendicular to Corridor 1. The lateral boundaries of all four corridors are demarcated by prominent moraines and hummocky terrain (Fig 5.1).

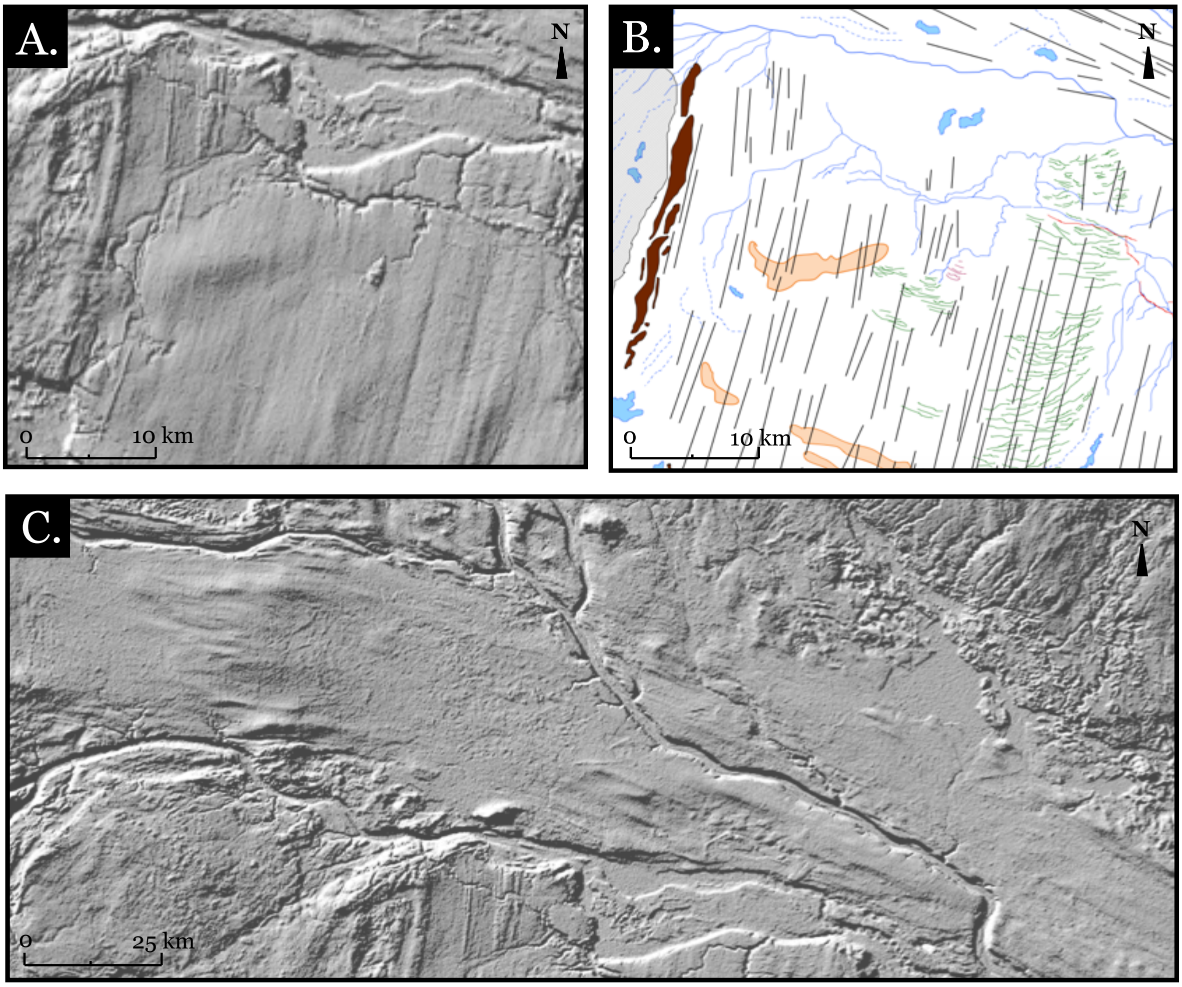


Figure 5.1: *SRTM imagery and resultant mapping of lineations along Corridor 1 and 2A.* ***A.*** *SRTM imagery of densely distributed, highly attenuated lineations at the centre of Corridor 1.* ***B.*** *Corresponding mapping of lineations in Corridor 1 (see Map Sheet 1 for full image). Lineations are drawn as black lines.* ***C.*** *SRTM imagery of smaller, less densely spaced lineations running northwest to southeast on the central section of Corridor 2A.*

Lineations, consisting of longitudinal ridges and furrows, within all corridors display high levels of spatial coherency (Fig 5.2). Lineations are most densely spaced within Corridors 1 and 2 A and B, where some are <150 m apart. The longest lineations are found within the zone of Corridor 1 bordered by Corridors 2A and B (Fig 5.2). Here lineations are >60 km long. In contrast features in Corridor 2C are not as densely grouped and are shorter, at 2-20 km long. There is a noticeable lack of lineations in the centre of this corridor, corresponding to a dense area of sand dunes (Great Sand Hills) (Map Sheet 1). Due to their subtlety these features could not be identified on Landsat ETM+ especially in areas where agricultural manipulation of the landscape has occurred. Thus mapping of lineations within this study was limited to SRTM imagery and only large lineations could be identified. It is acknowledged that much smaller lineations may exist that have not been mapped at this scale. Because of the resolution of the SRTM imagery, on which large streamlined features were identified, elongation ratios (ERs) could not be accurately determined, however, the majority of lineations have ERs of >10: 1.

5.1.2 Transverse Ridges (Types 1-4)

Three types of transverse ridges are identified within the SWSS. These ridges are so called as they are aligned transverse to the predominant orientations of lineations in each of the four corridors identified. These features are most commonly non-streamlined though in the centre of some corridors ridges have been smoothed and decorated with flutings indicative of streamlining. Each of these transverse ridge types are classified based on their characteristics described below.

5.1.2.1 Transverse Ridge: Type 1

Clearly visible on the SRTM imagery (Fig 5.2) are numerous type 1 transverse ridges (TR-1), which mainly occur as densely spaced arcuate ridges that collectively display lobate planforms. These ridges are split into two categorises based on size. Minor TR-1, mapped as lines on Figure 5.1 and 5.2 (predominantly from Landsat ETM+ imagery) are typically 1-10 m high, 50-100 m wide and up to 2 km long. Major moraine ridges in contrast are much larger (visible on both Landsat ETM+ and SRTM imagery), typically 1-3 km wide and can be up to 60 km long and 50 m high, thus their outlines are clearly visible and they have been mapped as polygons (Fig 5.2). A total of 154 TR-1 are mapped in the study area of which ~40 % have been previously identified (Campbell, 1986a, b, 1987a, b; Ross et al., 2009; Ó Cofaigh et al., 2010).

The largest series of major TR-1 occurs at the southern end of Corridor 1 (orientated perpendicular to associated lineations) and coincides with an abrupt rise in topography of the Cypress Hills. Individual TR-1 segments that make up this arc range in length from 1-60 km and in width from 0.5-2.5 km. They have been accentuated in the south due to the incision of meltwater channels. This series is broadly arcuate and extends into southern Alberta. Several prominent major TR-1 also occur, oriented parallel to their corresponding corridors' long axes, and mark the division between each smoothed corridor and the surrounding hummocky terrain. The largest of these are mapped along the western edge of Corridor 1, forming a 70 km discontinuous ridge. In profile these features comprise a single asymmetric ridge a few kilometres long and 20-30 m wide. Additionally along the northwest-southeast trending Corridors 2A and B, concentrations of minor TR-1 are present in several well preserved concentrations that have an inset lobate form that follows the local contours of the topography.

5.1.2.2 Transverse Ridge: Type 2

Type 2 transverse ridges (TR-2) have been mapped as two networks within the study area (Fig 5.3). Set 1 is located within Corridor 1. The network is fragmented into 3 sections separated by Corridors 2A and B, however in total the network stretches for ~300 km. Ridges are thin, sharp crested and extend on average between 100-400 m, although a few individual ridges extend to 5000 m in length (Fig 5.3). The interval composition of these ridges is reported to be almost entirely composed of diamicton based upon borehole records accessed for this within this study and internal analysis of the features by Campbell (1986a, b, 1987a, b). The second set was identified in the centre of Corridor 2A. This much smaller network stretches for just under 10 km, with ridge length varying between 400 m and 100 m. Morphologically they are very similar to set 1 and are characterised by very thin, low amplitude, sharp crested ridges. Measurements of the orientation of individual TR-2 taken from the SRTM imagery (Map Sheet 1) reveal a prominent WNW-ESE alignment with a subordinate WSW-ENE alignment for set 1 and a NNE-SSW alignment for set 2.

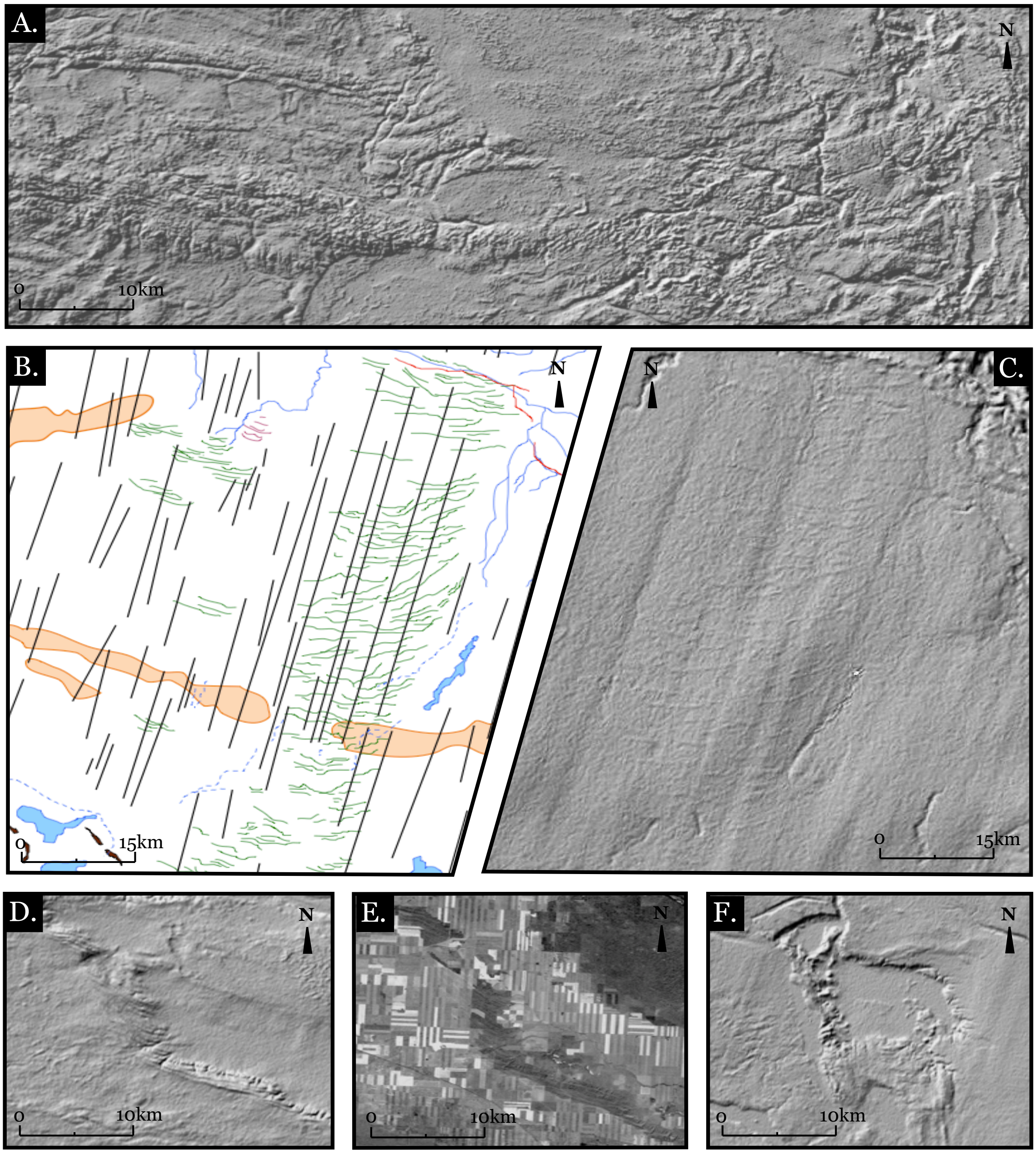


Figure 5.2: *SRTM, Landsat ETM+ and geomorphological mapping of transverse ridges types 1-3.* ***A.*** *SRTM imagery of TR-1. Note the densely spaced nature and arcuate planform of these ridges.* ***B/C.*** *SRTM imagery and geomorphic mapping of TR-2 (concentration 1). Ridges mapped as green non-linear lines (see Map Sheet 1 for full image).* ***D, E.*** *SRTM and Landsat ETM+ imagery of a TR-3 ridge complex in the southern section of Corridor 1.* ***F.*** *TR-3 complex in centre of Corridor 1 associated with a small depression.*

5.1.2.3 Transverse Ridge: Type 3

Large subparallel, generally arcuate, sharp crested ridges 10-60 m high and 150-300 m wide are visible throughout the study area. In many cases these ridges are accompanied by a depression (typically filled with a lake) immediately behind the end of the ridge arc, such as can been seen in front of Muddy Lake (40 km west of Handel) (see Map Sheet 1 for location)

5.1.2.4 Transverse Ridge: Type 4

Several large, wide and low amplitude ridge complexes (10 m high and 20 km wide) are visible in the northern portion of the study area that have been glacially smoothed by overriding ice. The ridges occur in two concentrations that form a southwesterly oriented arc. All type 4 ridge (TR-4) complexes are overprinted with lineations and their tops smoothed extensively. Consequently, individual ridge crests are not distinguishable and thus these smoothed mounds are mapped collectively as ridge complexes.

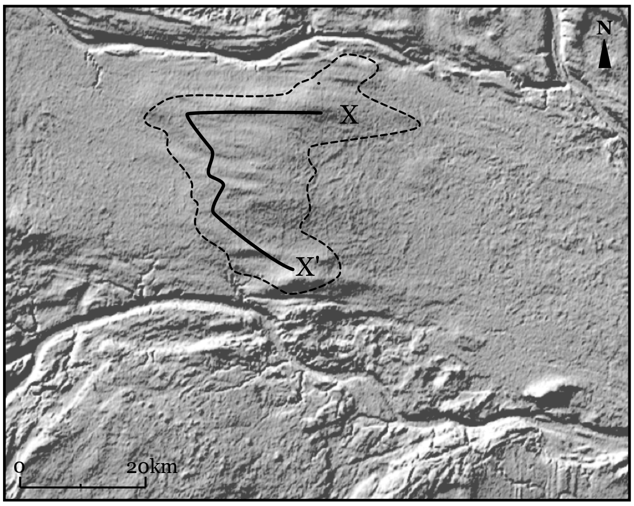


Figure 5.3: *SRTM of transverse ridges (type 4) on Corridor 2A, which have been overprinted by readvancing ice. Note the lineations that run southeast across this ridge complex. The extent of this landform is marked by points X and X' , its planform is represented by a solid black line and its extent by a dashed line.*

5.1.3 Inter-corridor and hummocky terrain

The large areas that lie between and thus delimit the margins of corridors of smoothed/streamlined terrain in the SWSS contain a range of glacial landforms with no consistent orientation (Map Sheet 1). The resolution of SRTM and Landsat ETM+ imagery is insufficient to provide significant insight into the morphology of this terrain.

5.1.4 Sinuous ridges

Elongate sinuous ridges are prominent throughout the study area and those mapped have heights between 5-10 m and lengths from 2-10 kms. These landforms were primarily identified from SRTM imagery thus only large lineations could be identified and it is acknowledged that much smaller lineations may exist that have not been mapped at this scale. While the majority of sinuous ridges remain isolated and relatively linear, within Corridor 1, a small complex of these ridges is present. These features are 10-15 km long and previously have been informally named the Glidden Esker Network (Christiansen, 1987). The sinuous ridges identified along Corridor 1 exhibit a northwest to southeast orientation following the natural slope. Further to the north on this corridor another sequence of prominent ridges are present. They form a 10 km network that occurs in close association with a collection of TR-3. Additionally, multiple sinuous ridges are also present within Corridors 2A, B and C and here exhibit a more west to east orientation (Map Sheet 1).

5.1.5 Abandoned channels

A complex and extensive system of abandoned channels occur throughout the study area. Abandoned channels are predominantly mapped from SRTM and from Landsat ETM+ data, however due to the small scale of many channels, where the resolution of imagery hindered identification, these channels were cross referenced against Campbells' (1986a, b, 1987a, b) glacial geology maps of the region. It should be noted that it is possible that some of the mapped channels are in fact sub-aerial rivers (seasonally active or dry), but it is difficult to conclusively differentiate between these using the imagery available. Within Corridor 1 abandoned channels predominantly have a NE-SW orientation. However within Corridors 2A, B and C, channels also exhibit a more easterly orientation. Abandoned channels are also common throughout the hummocky terrain sections, often occurring as a complex network of interconnected channels. In this terrain channel orientation is often chaotic.

5.2 Quaternary and glacial stratigraphy

17 stratigraphic units are recognised within the study area; seven that contain glacial diamicton and ten that do not. As outlined in the methods section of this thesis, units are differentiated on the basis of stratigraphic position, grain size, lithic composition and geophysical properties. In order to aid interpretation and to enable comparison, descriptions are summarised in Table 1. The inclusive stratigraphic model is displayed in Figure 5.4 and is also viewable as a dynamic 3D image on the CD-ROM in the inside back cover of this thesis (Appendix 2). For each unit included, a description of the unit, its thickness and distribution, how it is differentiated from other units and the nature of its contacts are described (Table 1). While not all units relate to terrestrial ice streams active during the Late Wisconsinan, a description of all units are given to aid in reconstructing the glacial history of the region and placing inferences about Late Wisconsinan ice streams into stratigraphic context.

By lithologic convention it is common practice within Saskatchewan and Alberta to refer to all 'stratified gravel, sand, silt and clay that overlies bedrock and underlies till' as the Empress Group (EMP) (Whitaker and Christiansen, 1972). This group is recognised within most of the major buried valleys in southwestern Saskatchewan and is inferred to be Quaternary pre-glacial, most likely Late Tertiary to early Quaternary in age (Whitaker and Christiansen, 1972; Evans and Campbell, 1995; Cummings et al., 2012). Within the study area, three separate lithologically distinct units can be mapped: unit 1, a basal sand and gravel containing clasts; unit 2, a middle silt and clay with minor sand and gravel beds; and unit 3, an upper glacial sand and gravel.

Directly overlying Quaternary pre-glacial sediments is Diamicton Unit 1 (DU-1). This clay rich diamicton is restricted within and along major buried valleys and channels. A thick sand rich layer, SU-A, then directly overlies DU-1, exhibiting a sharp contact with the overlying Diamicton Unit 2 (DU-2). This diamicton is widespread and differentiated from the overlying Diamicton Unit 3 (DU-3) only where a thin layer of stratified clay, sand and gravel is present (SU-B). Stratified Unit C (SU-C), composed of silt and clay, separates DU-3 from the overlying Diamicton Unit 4 (DU-4). This diamicton is present only in the north of the study area; thus SU-C is often overlain directly by Diamicton Unit 5 (DU-5), a thick, silty-sand rich diamicton. Stratified Unit E (SU-E) separates DU-5 from the overlying clayey Diamicton Unit 6 (DU-6); this unit is widespread and commonly contains incorporated masses of displaced older sediment. DU-6 is overlain by two different units. In three low topographic areas DU-6 is overlain by a thin layer of diamicton, Diamicton Unit 7 (DU-7) (separated in >20 boreholes by a very thin (<4 m) Stratified Unit F (SU-F)). However in the majority of the study area DU-6 is directly overlain by post-glacial, surficial stratified deposits (SSD).

Table 5.1. *Summary of subsurface-stratigraphy of the SWSS.*

Unit	No. of samples	Unit description	Maximum Thickness (m)	Elevation Range (m a.s.l.)	Origin
Surficial Stratified Deposits (SSD)	104/197	*Stratified gravel, sand, silt and clay*	*76*	*469-799*	*Fluvial, lacustrine or aeolian origin*
Diamicton Unit 7 (DU-7)	59/197	*Diamicton; silty-sand*	*20*	*782-467*	*Till*
Stratified Unit F (SU-F)	20/197	*Stratified sand, gravel and silt*	*4*	*707-511*	*Glaciofluvial or glaciolacustrine origin*
Diamicton Unit 6 (DU-6)	163/197	*Clayey and sandy diamicton, commonly containing incorporated masses of glacially displaced sediment*	*30*	*796-504*	*Till*
Stratified Unit E (SU-E)	79/197	*Stratified sand and gravel and small amounts of silt and clay*	*37*	*779-494*	*Glaciolacustrine or glaciofluvial origin*
Diamicton Unit 5 (DU-5)	136/197	*Silty-sand rich diamicton; very coarse sand rich in carbonate fragments*	*41*	*799-468*	*Till*
Stratified Unit D (SU-D)	35/197	*Stratified sand and gravel*	*13*	*737-490*	*Glaciofluvial origin*
Diamicton Unit 4 (DU-4)	47/197	*Clayey glacial diamicton*	*46*	*730-476*	*Till*
Stratified Unit C (SU-C)	62/197	*Stratified silt and clay in many locations also contains small amounts of sand and gravel*	*38*	*745-467*	*Glaciolacustrine origin*
Diamicton Unit 3 (DU-3)	117/197	*Sand rich glacial diamicton*	*66*	*743-456*	*Till*
Stratified Unit B (SU-B)	16/197	*Stratified clay, sand and gravel*	*24*	*683-559*	*Glaciofluvial origin*
Diamicton Unit 2 (DU-2)	61/197	*Clay rich glacial diamicton*	*52*	*745-451*	*Till*
Stratified Unit A (SU-A)	51/197	*Silt, sand and gravel*	*97*	*735-473*	*Glaciofluvial origin*
Diamicton Unit 1 (DU-1)	56/197	*Clayey glacial diamicton interbedded with well sorted clays, silts and sands, overlain by stratified sediment in some places*	*58*	*726-422*	*Till*
Empress (Unit 3) (EMP-3)	16/197	*Stratified sand and gravel; contains clasts derived from Canadian Shield*	*35*	*706-453*	*Glaciofluvial origin*
Empress (Unit 2) (EMP-2)	47/197	*Stratified clay and silt*	*49*	*699-451*	*Lacustrine or fluvial origin*
Empress (Unit 1) (EMP-1)	42/197	*Stratified sand and gravel, mainly chert and quartzite derived from the Cordilleran*	*70*	*680-415*	*Preglacial fluvial origin*
Bedrock	164/197	*Defined as all lithified clastic sediment that underlies the early Tertiary erosional surface commonly of crystalline Precambrian rock overlain by sedimentary rock cover*	*N/A*	*N/A*	*N/A*

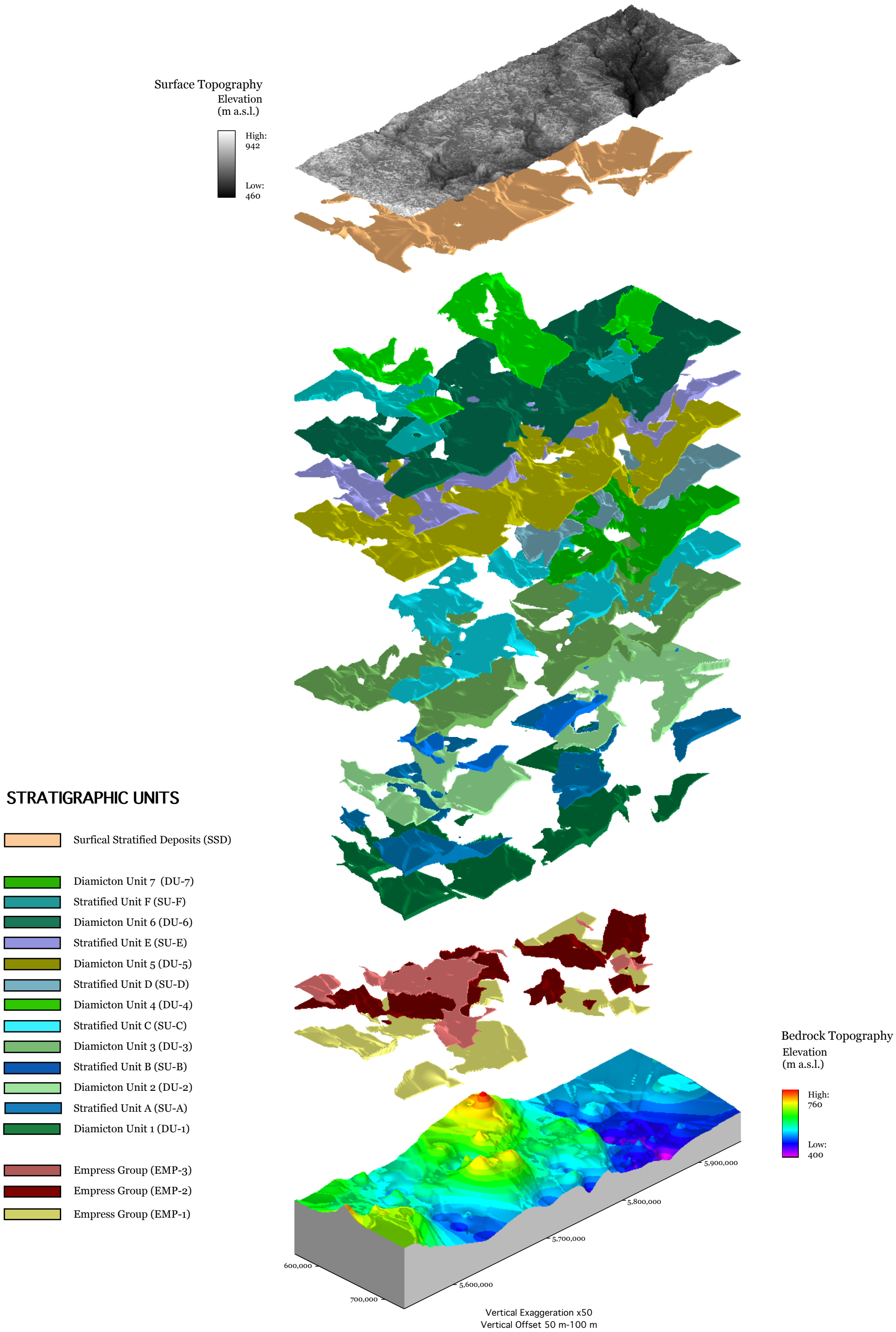


Figure 5.4: *Exploded stratigraphic model of all 17 stratigraphic units recorded within the SWSS. Model is overlain with a SRTM DEM of the study area, and underlain with a bedrock topography map (see Section 4.2.5 for details of production of bedrock topography maps). This diagram is also viewable as a dynamic 3D image provided on a CD in the inside back cover of this thesis.*

5.2.1 Bedrock surface topography: palaeo valley systems

The close relationship between the bedrock topography and physiography of the land surface in the SWSS indicates that large parts of the contemporary landscape reflect the topography of the underlying bedrock (Map Sheet 1 and 2). Thus examination of the underlying topography is necessary to aid understanding of landscape evolution and dynamic ice stream behaviour. In this study the term 'bedrock' is defined as any lithified clastic sediment lying below the early Tertiary erosional surface (Andriashek and Fenton, 1989). The bedrock within the study area is composed of crystalline Precambrian rock, which is overlain by a sedimentary rock cover (Cummings et al., 2012). Within boreholes this sedimentary rock cover is predominantly poorly consolidated, carbonate poor, cretaceous shale. Incised into this bedrock surface are multiple preglacial valleys, which have been subdivided into discrete systems. Two major preglacial valley systems (Tyner and Battleford) have been previously named and inferred on a regional scale (Stalker, 1961).

Battleford Valley and tributaries

The Battleford Valley system is the most extensive valley in the study area and is likely the preglacial valley of the ancestral North Saskatchewan River. It enters the northwest corner of the study area above Manitou Lake as three tributaries called Lloydminster and Wainwright channels and North Battleford Valley (names assigned based on closest settlements), all of which trend southeast. Based on borehole measurements, the contour lines in Figure 5.5 show the southeast segment of the North Battleford Valley slope sharply southeast. A southeastern slope is also assumed for the northwestern section of this valley, however, due to the low resolution of cores in this area, it is difficult to accurately define the valley depth and width.

Tyner Valley and tributaries

The main valley enters the study area, in the extreme southwest, and extends to the east. Compared to the Battleford Valley system, the Tyner Valley system is wider with more gentle sloping walls. The valley follows a similar path to the present day Saskatchewan River and is likely the ancestral preglacial valley of the Saskatchewan River. However, borehole records within the channels show the presence of glaciofluvial sediment. This indicates that parts of this valley were occupied by glacial meltwater and may have been subsequently altered by that meltwater during the first glaciation of the region (see Section 7.1).

5.2.1 Empress Group

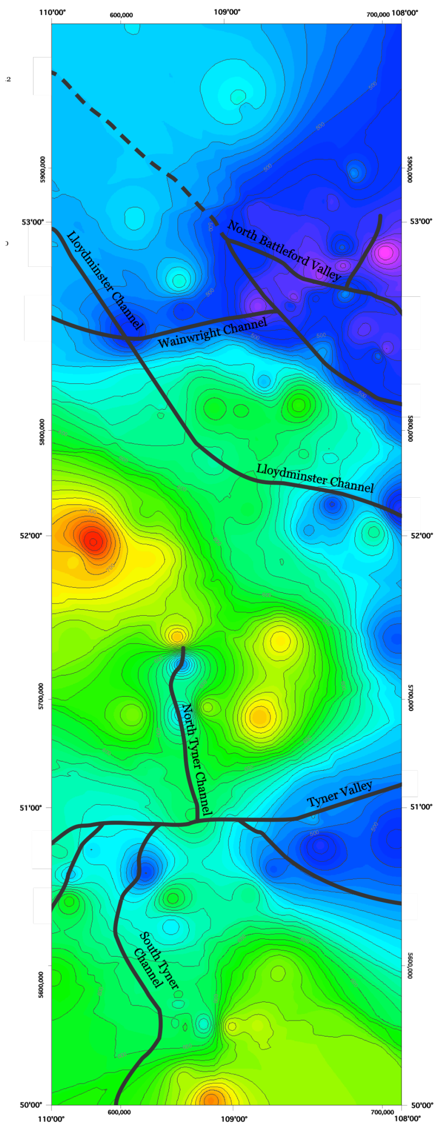
Unit 1: Preglacial sand and gravel

Unit description

In general the unit consists of <20 m of basal gravel overlain by sand and gravelly sand, which ranges from 5-40 m in thickness (Fig 5.6). Sands and gravels that dominate the composition of unit 1 of the Empress Group (EMP-1) are found on the floor of the Tyner and Battleford buried valley systems. Drillers' logs indicate that the unit is composed of clasts derived from either local sandstone or the Cordillera. The lower gravel layer is predominantly light-coloured quartzite. Similarly the sand is light coloured quartz and dark coloured chert. This mixture of chert, quartz and quartzite gives the unit a 'salt and pepper' appearance, noted in drillers' log descriptions. The sand is generally medium-grained and well sorted and is unoxidised in all cores with the exception of BH-033. In several locations within the eastern part of the North Battleford Valley 'till-like' deposits are recorded at the base of the Empress Group beneath gravel, though such deposits are interpreted as slump deposits on drillers' logs.

Thickness and distribution

In general deposits of EMP-1 lie within the Battleford and Tyner buried valley systems (Fig 5.7). The unit is not recorded on the interfluves between these valleys. Thus the elevation of this unit is restricted to between 680 m (along the very edge of the Cypress Hills), and 415 m (along the North Battleford Valley). Within the eastern end of the Battleford and Tyner buried valley systems the surface of the unit slopes to the southwestward, indicating that the last stage of drainage in these valleys was in this direction (Fig 5.7), consistent with the northwestern slope of the region's topography.



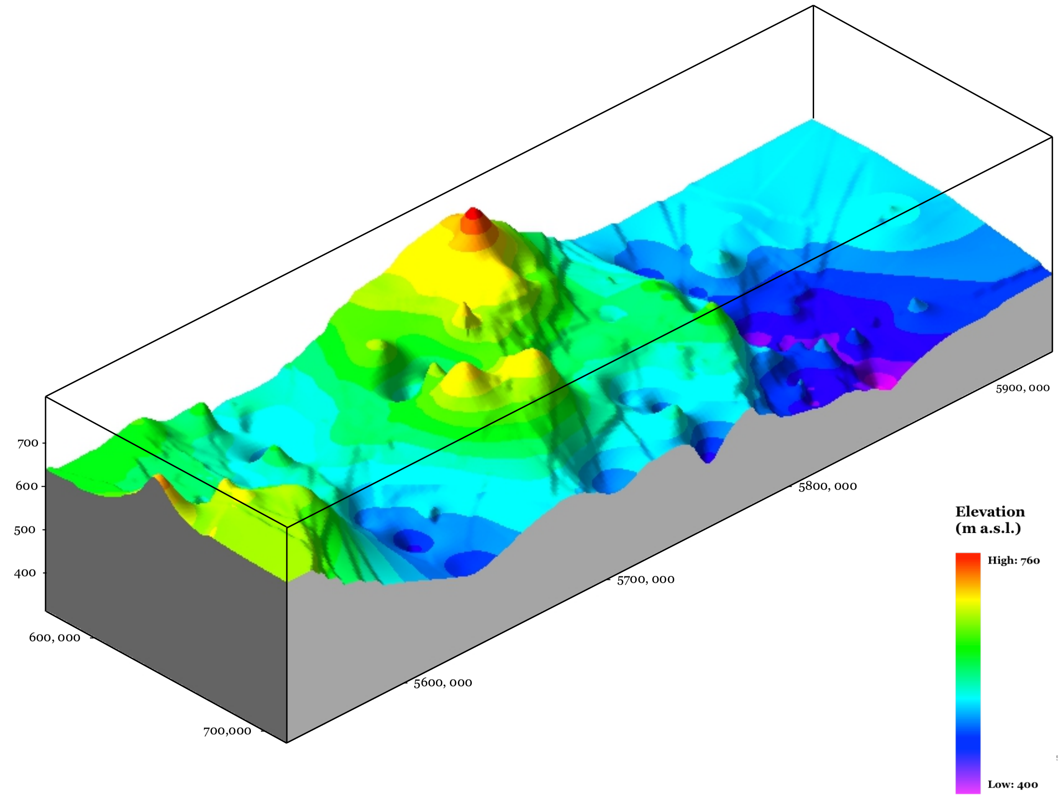


Figure 5.5: *Bedrock topography of the SWSS. Black lines denote buried valleys and channels. In areas where bedrock topography is accurately constrained (see Section 5.2.1 for details) paths are adjusted and additional channels added from those originally proposed by Stalker (1961). Black dashed lines indicate areas where the path of channels cannot be reliably constrained due to the low resolution of bedrock topography modelling, resulting from limited boreholes in some areas. (3D vertical exaggeration 50x)*

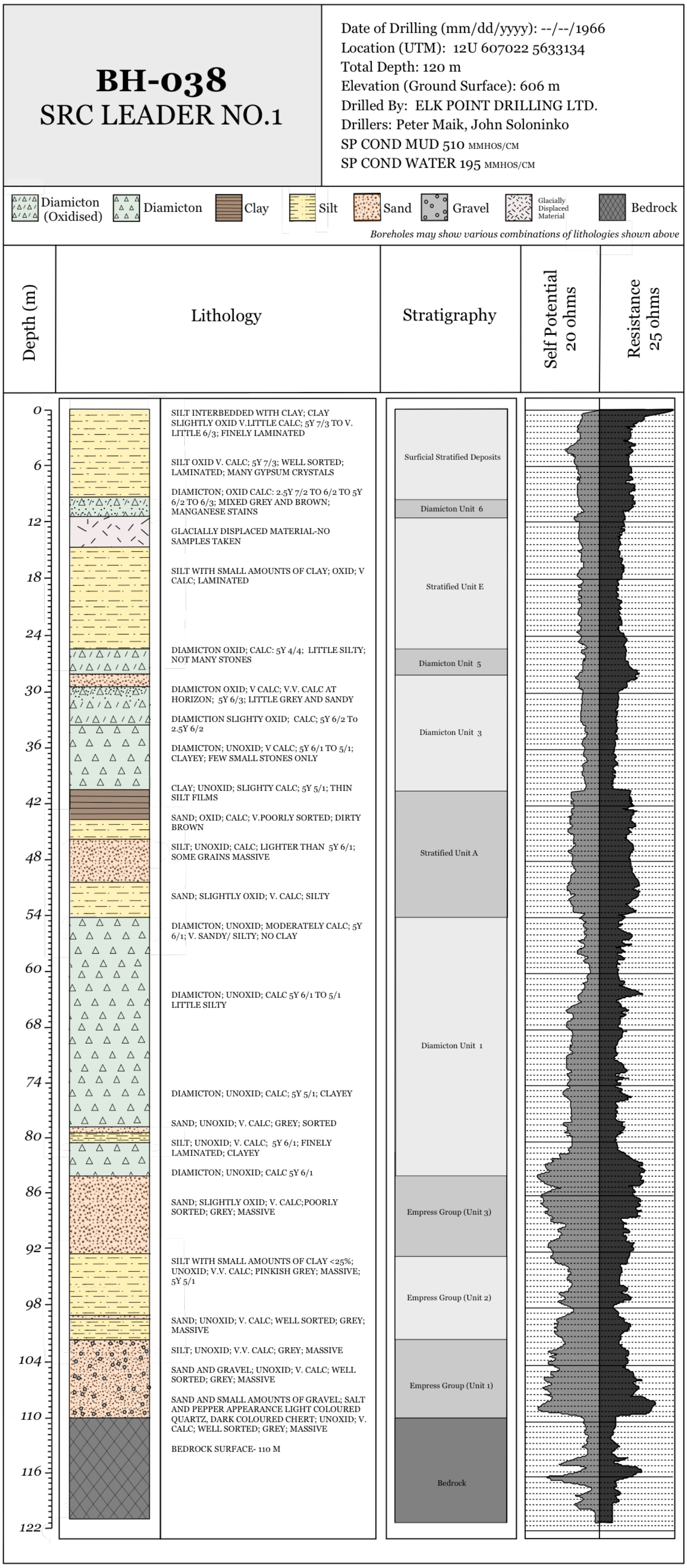


Figure 5.6: *Composite borehole log (Borehole ID-038) from the SWSS. Note the distinct change in electric log response between EMP-1 and EMP-2.*

Differentiation from other units

Differentiation of this unit from the underlying bedrock can be easily made based upon differences in grain size and resistivity response (Fig 5.6). This is particularly true where sand and gravel of EMP-1 are overlain by clayey deposits of EMP-2, and the two are easily differentiated by their electric-log responses. In contrast, where EMP-1 sands and gravels are directly overlain by glacial sand and gravel of EMP-3 the two can only be differentiated based on their elevation.

Nature of contact

EMP-1 shares an unconformable contact with the underlying bedrock and is easily recognised on electric logs (Fig 5.6). The upper contact of EMP-1 differs spatially within the study area. In segments of the Tyner and Battleford buried valley systems where EMP-1 is overlain by clays of EMP-2 or clay-rich diamicton (DU-1) the contact is sharp and easily recognised. However in places where it is overlain by EMP-3 (especially in the northern tributary of the Tyner Valley), the contact cannot be easily recognised and can only be estimated based on stratigraphic position and elevation range.

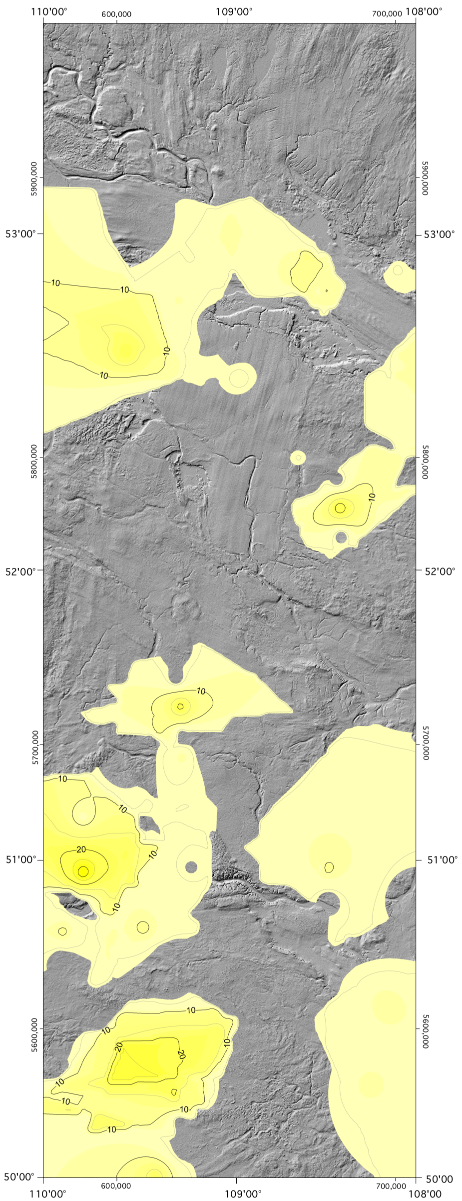
Unit 2: Silt and clay, minor sand and gravel

Unit description

Unit 2 of the Empress Group (EMP-2) is composed of grey (5Y 6/1 to 7/1) laminated clay and silt. Within the Battleford Valley system, sand and gravel are commonly interbedded with the clay.

Thickness and distribution

The unit is confined almost entirely to the bottom of the Battleford and Tyner buried valley systems and thus ranges in elevation from 699 m to 415 m (Fig 5.8). Along the Battleford Valley system, deposits of EMP-2 range from 2-46 m in thickness. Within the main Tyner Valley, EMP-2 is in contrast thinner and more sporadic in appearance (2-31 m). However the unit is considerably thicker within the northern tributary of the Tyner Valley, where silt and clay thicken to 70 m.



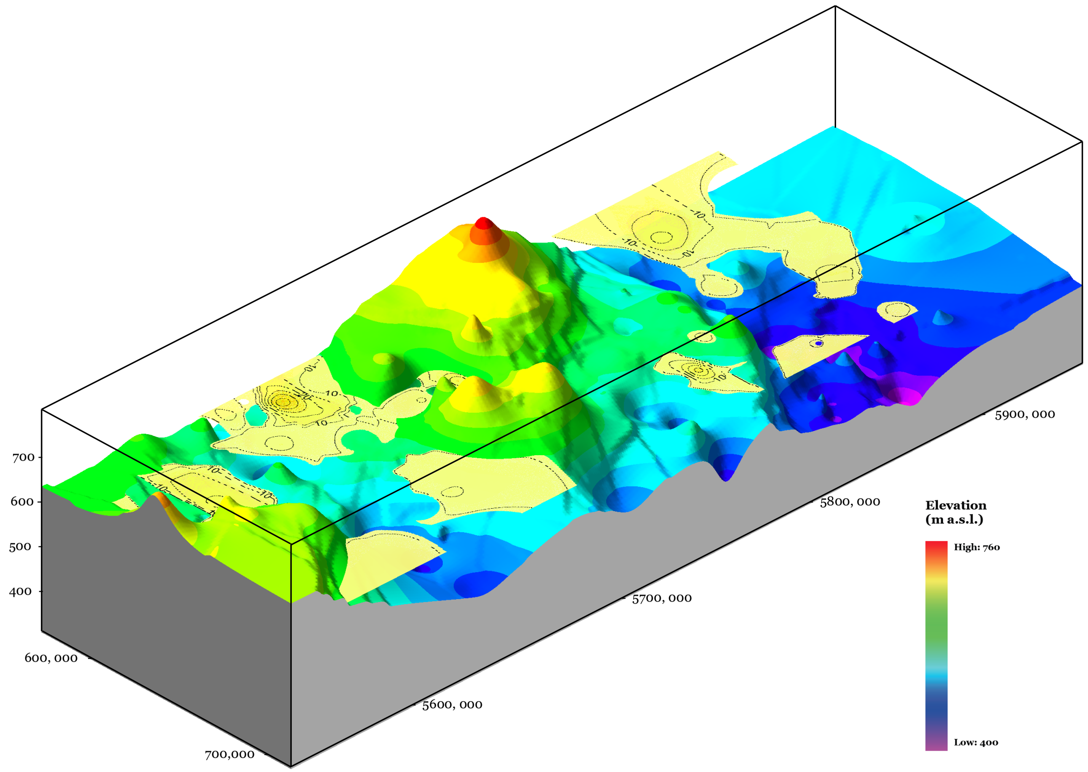


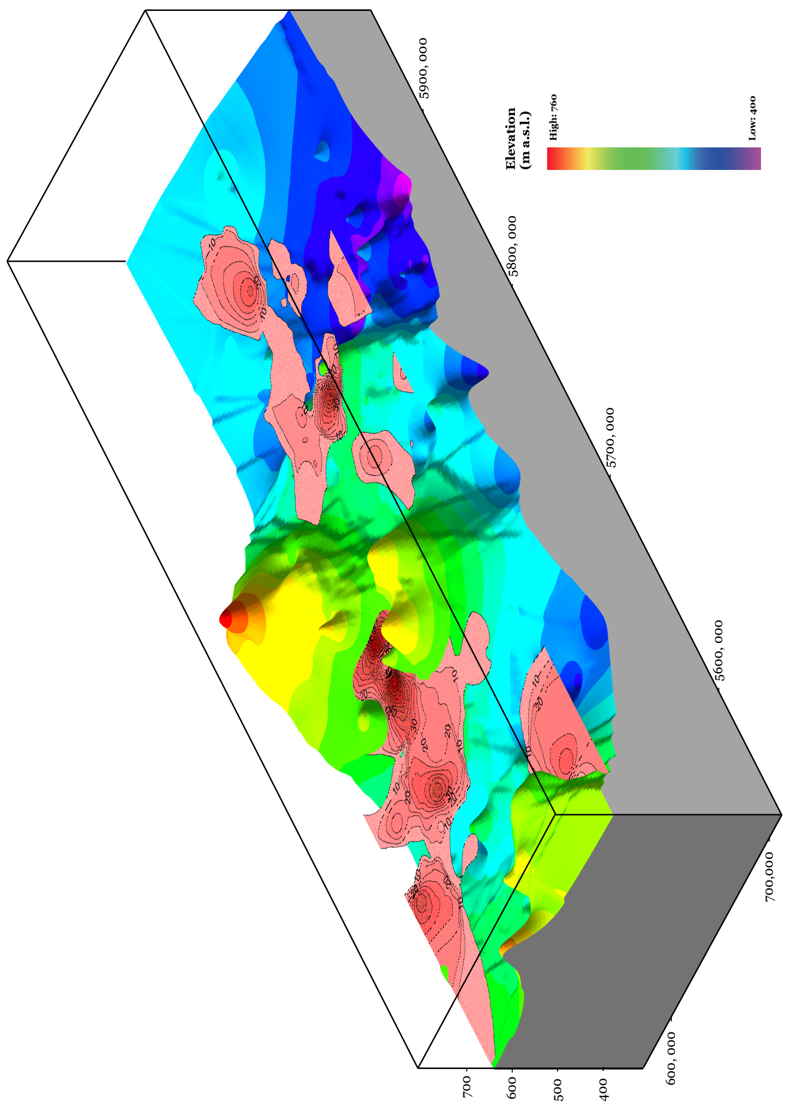
Figure 5.7: *2D and 3D representation of the thickness of sand and gravel comprising EMP-1. Note the distribution of this unit in relation to topographic lows (major buried valleys and channels). Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within Rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

Differentiation from other units

In areas where EMP-2 overlies sand and gravel of EMP-1 and is overlain by EMP-3 (composed of sand and gravel) it is identified with confidence. However, where this unit directly overlies bedrock, EMP-2 can be difficult to differentiate from the shale. EMP-2 silt and clay is differentiated from younger clayey units (SU-A to F), primarily based on its stratigraphic position, though it should be noted that this unit comprises a very similar composition and further laboratory analysis of sediments would be needed to differentiate this unit with complete confidence.

Nature of contact

The contact between EMP-1 and EMP-2 is discussed in relation to EMP-1 in the previous section. In the central Tyner Valley and segments of its southern tributary, EMP-2 directly overlies shale bedrock and this contact is challenging to recognise from borehole logs. In a few places where EMP-2 is overlain by till, the contact is easily recognisable from borehole descriptions. EMP-2 is overlain by sand and gravel of EMP-3 in several segments of the study area and this contact is generally sharp.



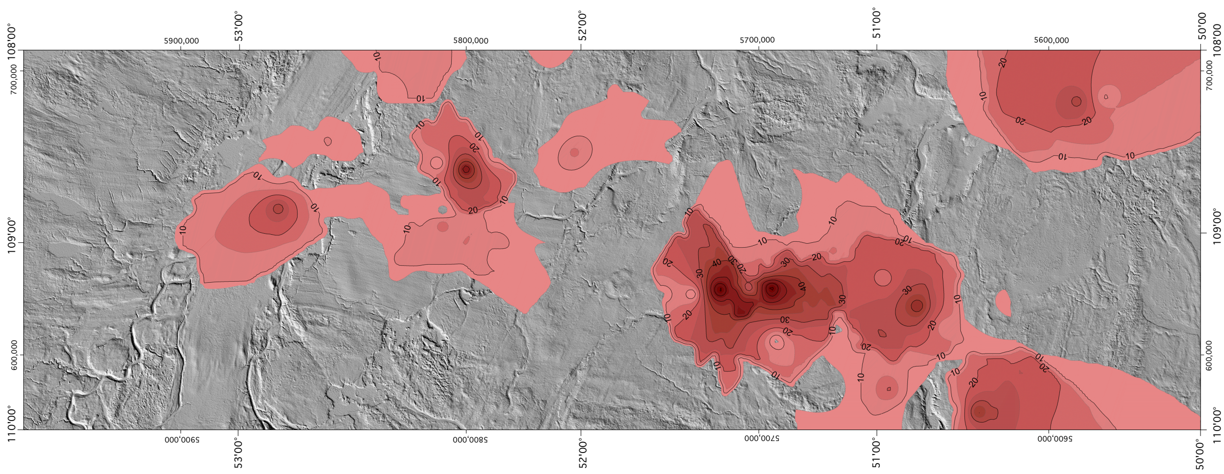


Figure 5.8: *2D and 3D representation of the thickness of silt and clay (minor sand and gravel) comprising EMP-2. Note the distribution of this unit in relation to topographic lows (major buried valleys and channels). Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within Rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

Unit 3: Glacial sand and gravel

Unit description

The composition of Unit 3 of the Empress Group (EMP-3) is dominated by sands and gravels with very small amounts of silt and clay also present. This unit is a very distinct pinkish brown when oxidised and pinkish grey colour when unoxidised (Fig 5.6). It is likely that the pink tinge is due to abundant potassium feldspar within the sand derived from granitic and metamorphic rocks. Gravels within this unit are typically composed of lithologies derived from the Precambrian Canadian Shield.

Thickness and distribution

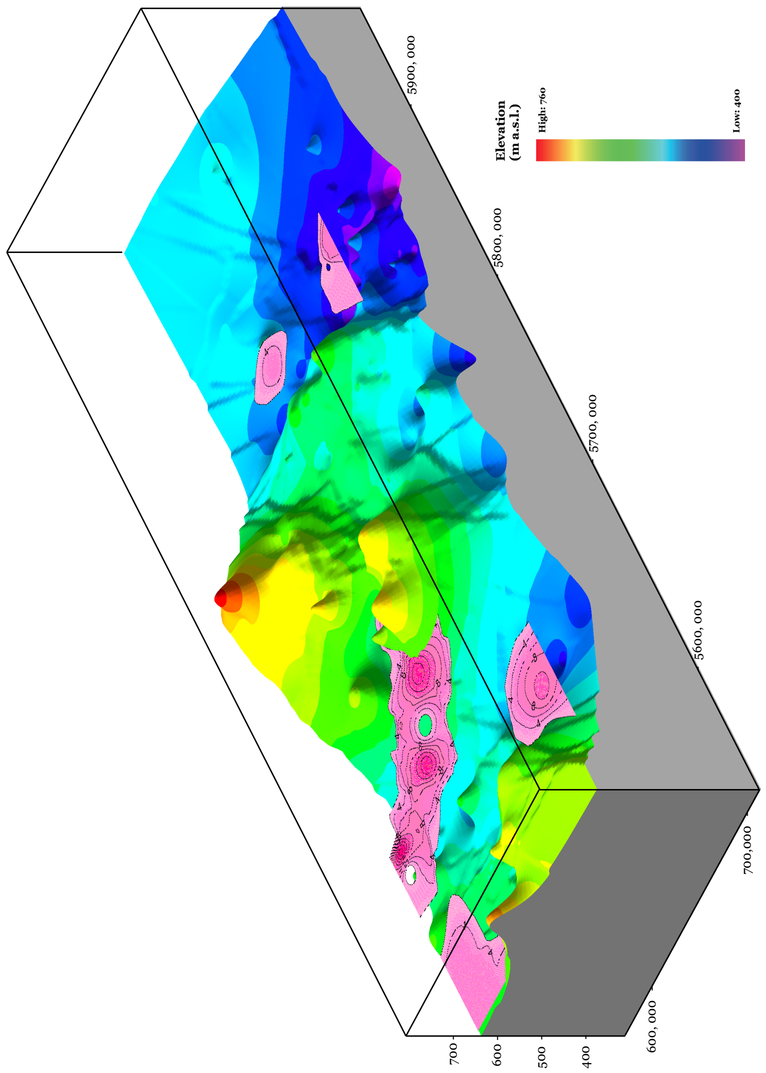
The extent of EMP-3 is more spatially restricted than the underlying EMP-1 and 2 and in the majority of locations lies within the base of the Tyner or Battleford buried valley systems. The top of the unit reaches an elevation of 706 m on the western Cypress Hills, and is recorded at 453 m at its lowest within the Battleford buried valley system (Fig 5.9).

Differentiation from other units

Recognition of EMP-3 is easiest when the unit is underlain by EMP-2 and overlain by DU-1, however this stratigraphic situation is only found in 9 cores in the North Tyner Channel. Thus in many places confident identification of this unit is difficult. Adding to this, EMP-3 can also be difficult to differentiate from younger sand and gravel deposits of SU-A if the two are in contact such as in BH-110.

Nature of contact

The contact between EMP-3 and the underlying bedrock or EMP-2 is easily recognised and is recorded as sharp. This contact is most easily identified by a sharp decrease in resistivity. The contact between EMP-3 and the overlying SU-A is however considerably more difficult to recognise. These two units have very similar composition and thus there is little to no difference in their log responses meaning they cannot be easily defined.



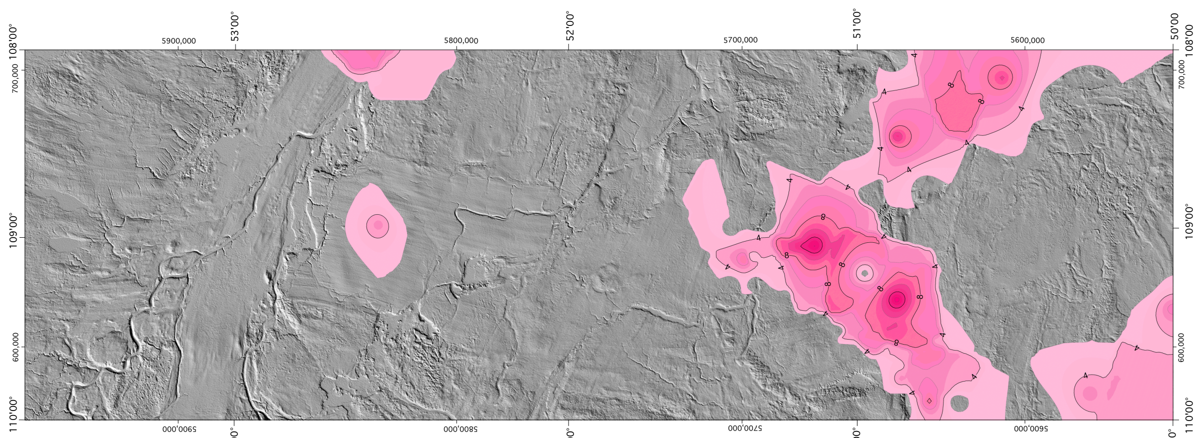


Figure 5.9: *2D and 3D representation of the thickness of sand and gravel comprising EMP-3. Note the distribution of this unit in relation to topographic lows (major buried valleys and channels). Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within Rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.2 Diamicton Unit 1

Unit description

This unit comprises clayey diamicton that in some places also includes sand (Fig 5.10). Using electric log responses, the unit is very easily recognised, especially where it lies between other stratified units. In the majority of boreholes this unit is predominantly unoxidised and is dark grey brown (2.5Y 4/2). Additionally the unit is interbedded with well-sorted sediment, predominantly clay but also silt and sand of undetermined origin. One such area is along the eastern end of the North Battleford Valley, where the unit consists of clay, silt and diamicton. In all areas the diamicton has very low resistivity, presumably the result of abundant clay.

Thickness and distribution

DU-1 lies primarily along major buried valleys and channels (Fig 5.11). The unit is also identified at the edges of some uplands, especially in the south of the study area. The thickness of the unit is generally <20 m. The unit is thickest (58 m) in the southwestern North Battleford Valley and the western end of the Tyner Valley (28 m).

Differentiation from other units

In places where this unit overlies coarse deposits of the Empress Group and where it is overlain by SU-A it is easily mapped (Fig 5.11). In places the unit directly overlies EMP-2 and is overlain by DU-2 (which is also clay rich) making differentiation more difficult. In such situations, differentiation is based on the change in resistivity and the lesser amount of clasts in DU-1, however due to the similarity it is possible DU-1 could be miscorrelated in some cases.

Nature of contact

In the majority of boreholes this unit overlies the Empress Group, and where the latter consists of sand and gravel the contact is sharp (Fig 5.10). However where the Empress Group consists of clay and silt it is difficult to establish the contact been the two units. This is exemplified within BH-144 and BH-004. Within these cores, drillers' logs record 'till/possible glacial till', which is interpreted as DU-1, even though electric logs show a minimal difference between this till and the underlying clay of

EMP-2. The upper contact between DU-1 and SU-A is interpreted as conformable and in most places is distinct and well defined on electric logs. However it should be noted that where the base of SU-A contains silt and clay the contact with DU-1 is not as sharp. An unconformable and generally sharp lower contact of DU-1 exists where this unit overlies bedrock within a few places at the very edges of uplands surrounding the Battleford and Tyner buried valley systems. Additionally in a number of boreholes (BH-005 and BH-038) drillers' logs describe inclusions of 'ice rafted shales' and 'claystone' within DU-1, which indicate that in places bedrock may have been plucked and then incorporated into DU-1 (Fig 5.6).

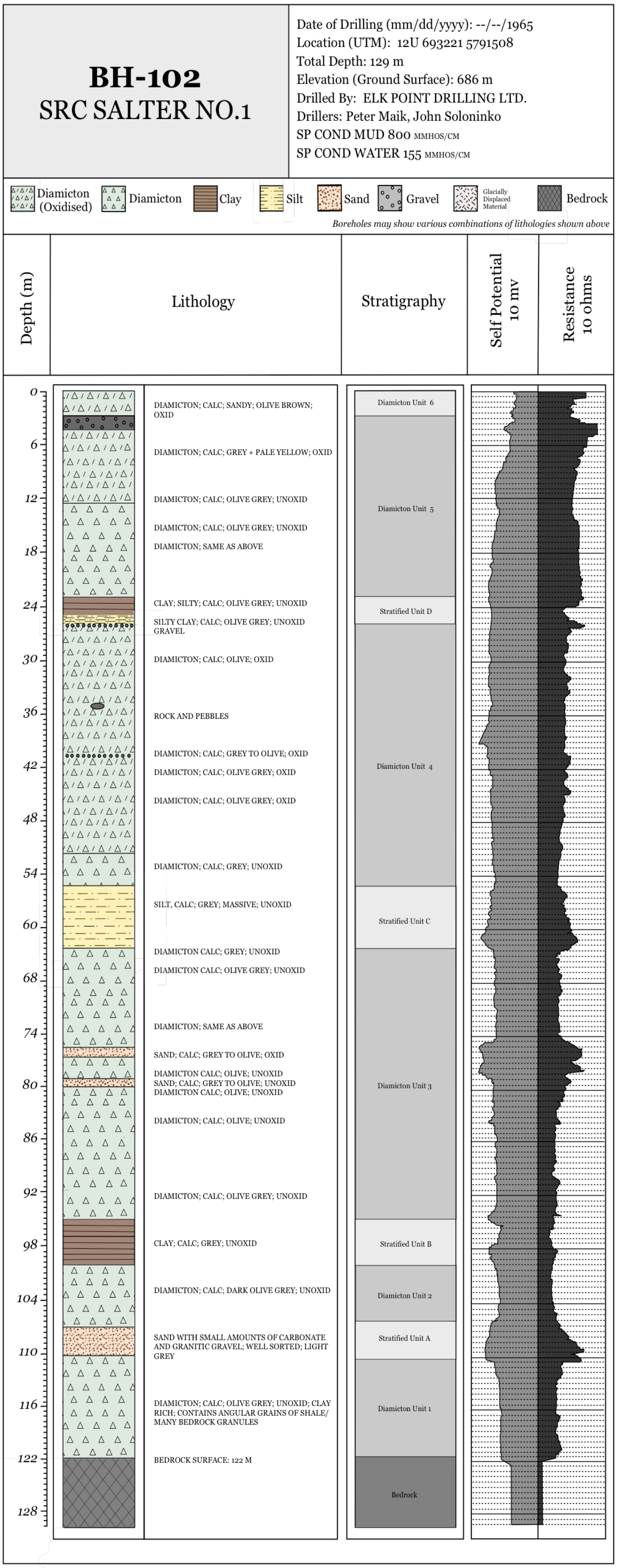
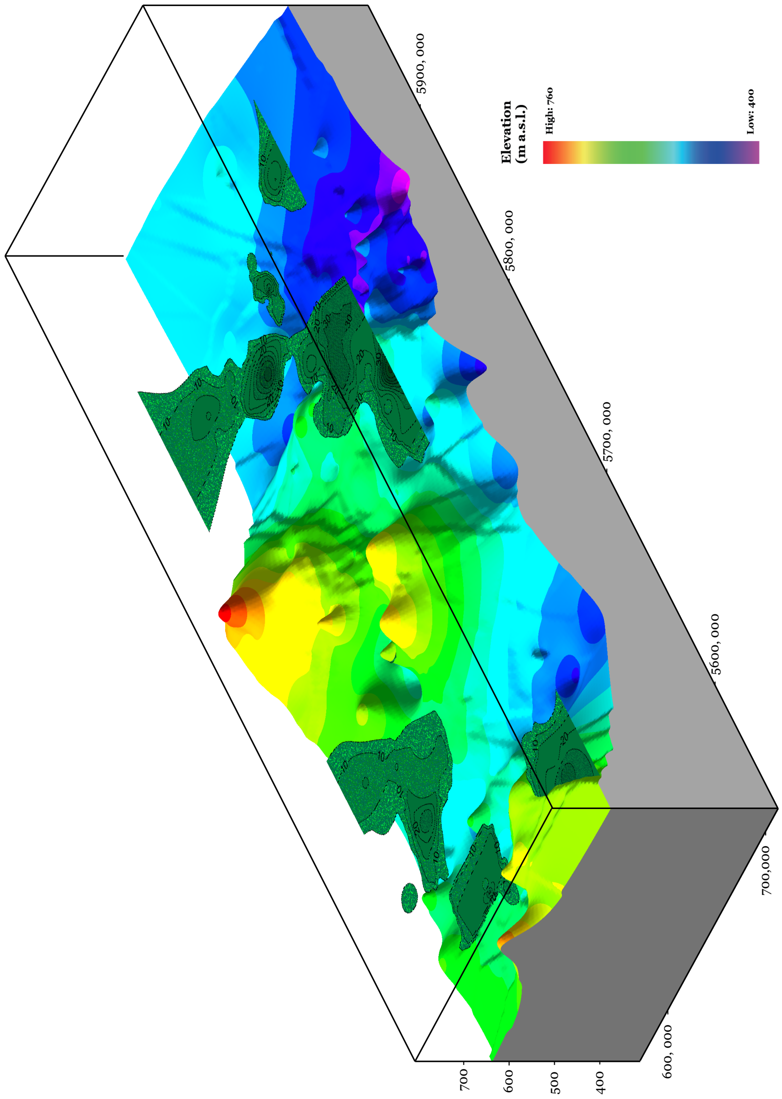


Figure 5.10: *Composite borehole log (Borehole ID-102) from the SWSS. Note the distinct change in electric log response between DU-1 and SU-A.*



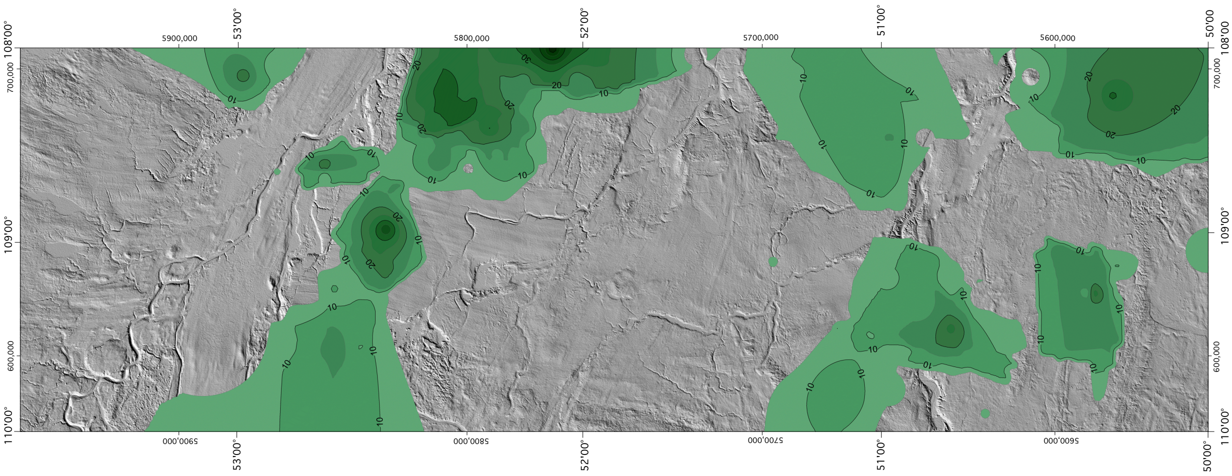


Figure 5.11: *2D and 3D representation of the thickness of clayey diamicton comprising DU-1. Note the distribution of this unit in relation to topographic lows (major buried valleys and channels). Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within Rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.3 Stratified Unit A

Unit description

The deposits of this unit are in general recorded as unoxidised and have a light to medium grey colour (Fig 5.10). This unit is made up of a majority of coarse grained sand and gravel but also contains minor amounts of silt and clay that are interbedded within the main sands and gravels.

Thickness and distribution

Within buried valleys and channels the unit overlies either DU-1 or Empress Group sediments. Its extent is larger than that of either the Empress Group or DU-1, thus in some areas it also overlies bedrock between some valleys. The distribution of this unit can thus be defined by two broad trends surrounding the Battleford and Tyner buried valley systems (Fig 5.12). Although, as discussed in section 5.2.1 data is lacking in the northwest of the North Battleford Valley, it is uncertain if the unit is present in this area. The top of unit SU-A ranges in elevation from as high as 580 m upon the Cypress Hills and is generally thick throughout the study area, ranging between 10-30 m, with the thickest deposits being in the eastern Tyner Valley and the southern tributary of the North Battleford Valley (Fig 5.12).

Differentiation from other units

Due to the composition of SU-A being very similar to other stratified units, stratigraphic position is the only way to differentiate them. In some places where SU-A is inferred to directly overlie the Empress Group differentiation between the two is based on the elevation of the deposits.

Nature of contact

Both the upper contact with DU-2 and the lower contact with DU-1 are sharp (Fig 5.13).

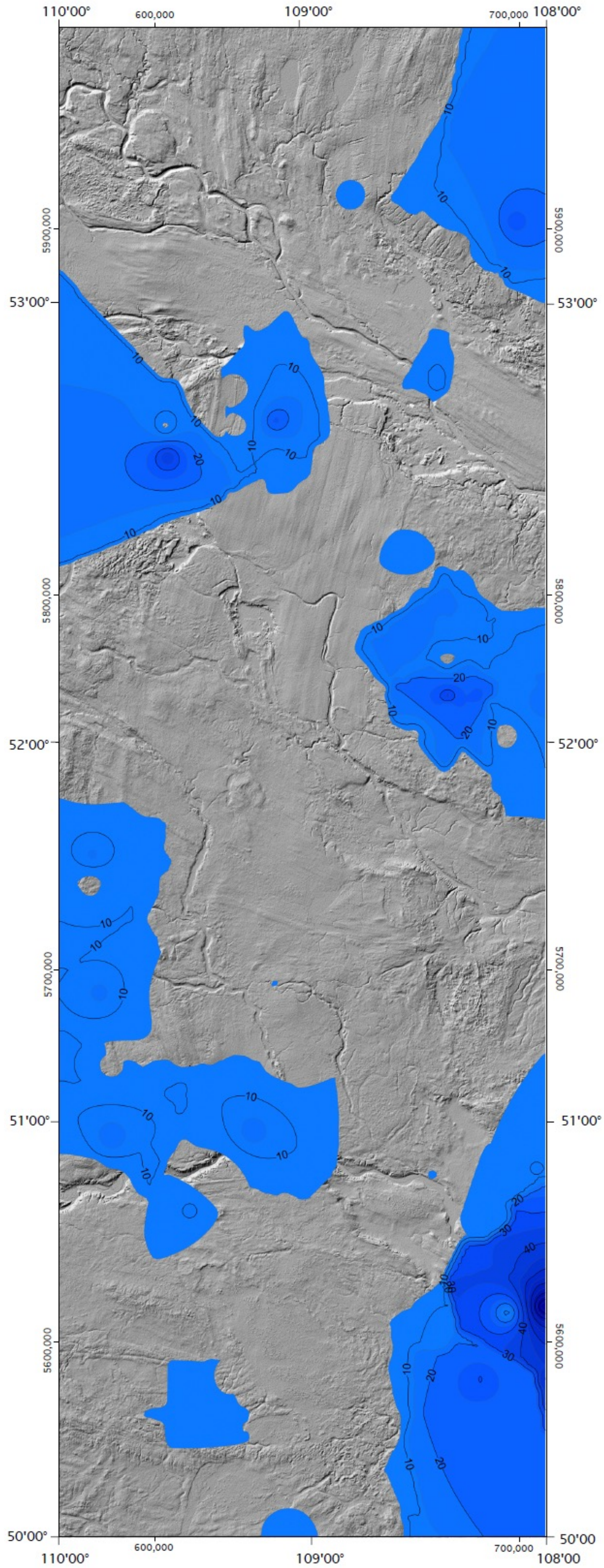


Figure 5.12: *Thickness of sand and gravel (with minor silt and clay beds) comprising SU-A. Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within Rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

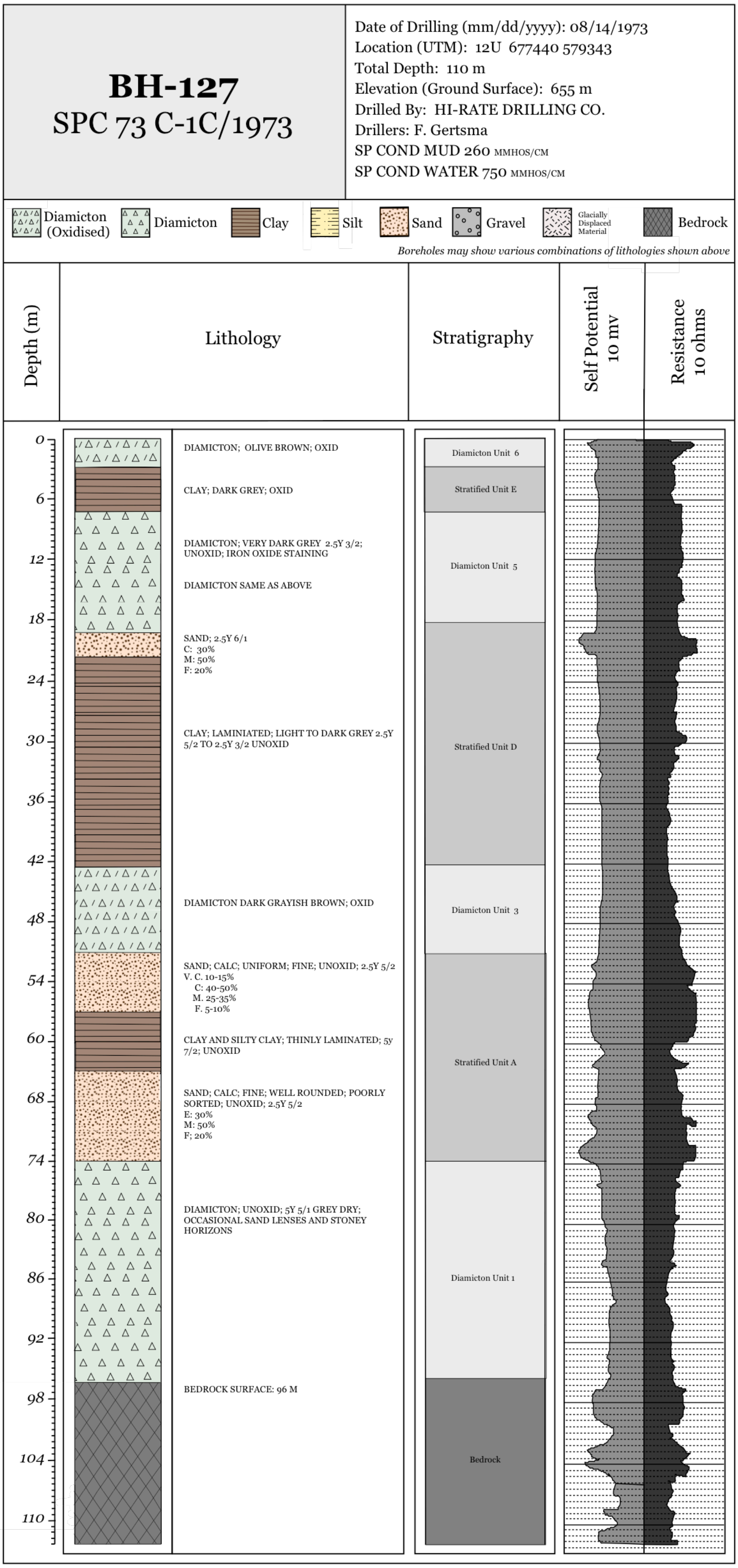


Figure 5.13: *Composite borehole log (Borehole ID-127) from the SWSS Saskatchewan. Note the low resistivity of DU-1.*

5.2.4 Diamicton Units 2 and 3

Unit description

DU-2 and 3 are very similar and can only be confidentially differentiated where a thin layer of stratified clay, sand and gravel (SU-B) is present, and in the vast majority of cases are only be differentiated based on slight differences in electric log response. Both units are very dark grey (5Y 3/1) when unoxidised and commonly olive brown (2.5Y 4/3) when oxidised. Due to the similarity in these units and the lack of SU-B in the majority of boreholes these units are discussed together.

Thickness and distribution

The majority of DU-2 is located within the Battleford and Tyner buried valley systems. The unit is also recorded within boreholes on the upland north of the Tyner Valley (Western Hills Upland) and the upland north of the North Battleford Valley (Turtlelake Upland) (Fig 5.14). The top of the unit ranges in elevation from 745 m in the Cypress Hills to 451 m in the Wainwright Channel and 449 m in the east Tyner Valley. The thickest deposits of this unit occur within the North Battleford Valley, where a localised thickening of the unit is present and as much as 64 m of DU-2 occurs and is overlain by 2-18 m of SU-B (Fig 5.14). DU-3 is more widespread than the underlying DU-2 and is only absent along a thin strip of upland centred on the Western Hills (Fig 5.14). The unit ranges in elevation between 743 m to 456 m and reaches a thickness of 69 m in the upland area north of the North Battleford Valley. Here the combined thickness of both units locally exceeds 100 m.

Differentiation from other units

Stratigraphic position is the only means by which position DU-2 can be differentiated from the underlying DU-1 and thus in many places the differentiation of these units is tentative. Furthermore DU-2 and 3 can only be confidently differentiated where SU-A lies between them, as both have similar grain sizes and electric log responses. The diamicton of DU-2 has a slightly lower resistivity than DU-3. Electric logs are also useful in differentiating DU-3 from the overlying SU-C or DU-4 where logs generally show a distinct break at the contact (Fig 5.10).

Nature of contact

The contact between DU-2 and 3 is generally sharp, though as discussed above, is in some places marked locally by SU-B, which separates the two units. In the majority of boreholes, DU-3 is overlain by SU-C and generally displays a sharp contact and can be easily recognised. DU-2 has a sharp and well-defined contact where it overlies SU-A. DU-2 also overlies bedrock in some locations, especially in the Western Hills Upland. Here the contact is gradational and difficult to recognise because DU-2 contains diamicton and gradually eroded bedrock material.

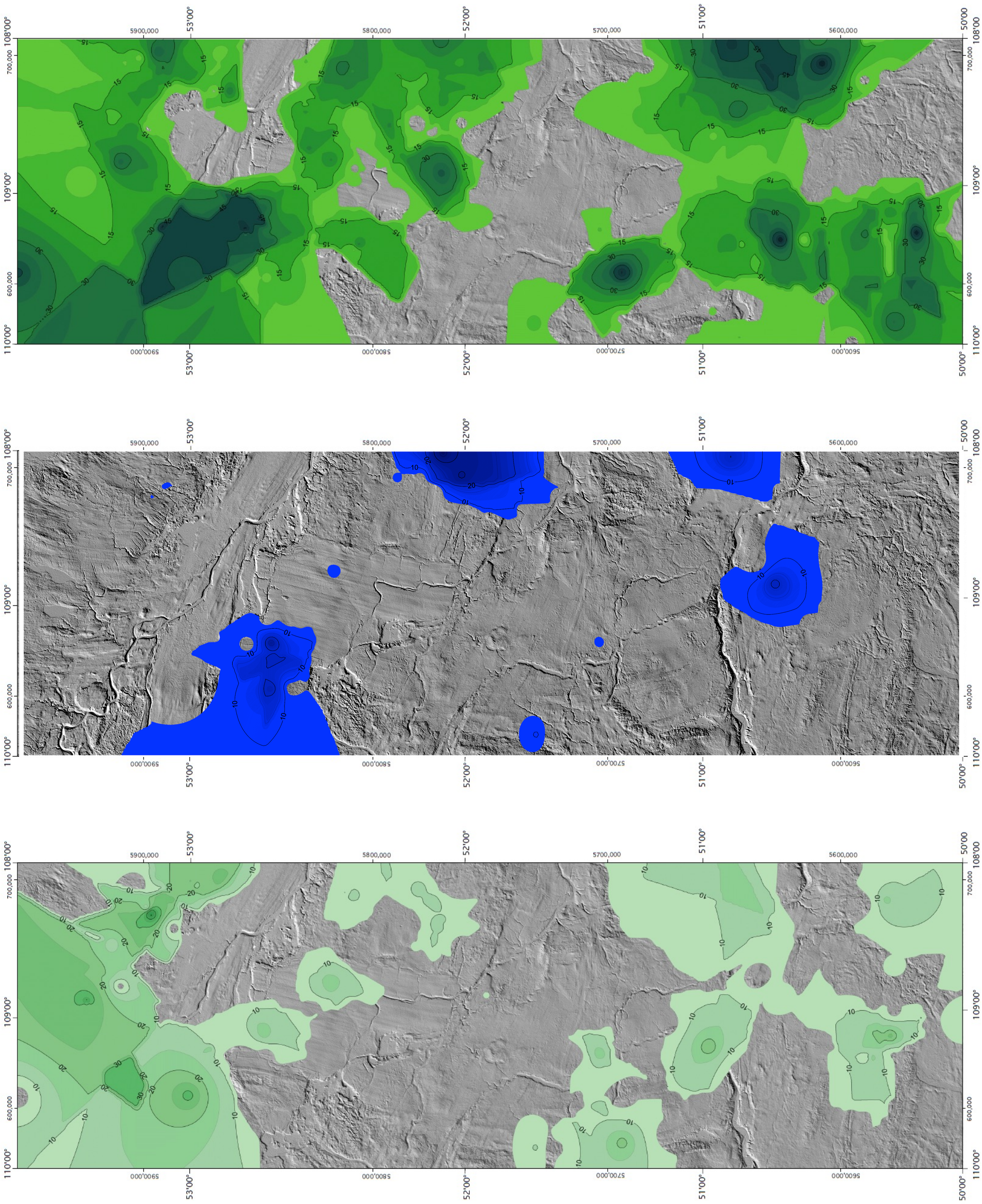


Figure 5.14: *Thickness of clayey diamicton comprising DU-2 (left); stratified clay, sand and gravel comprising SU-B (centre) and sandy diamicton comprising DU-3 (right). Due to the irregular distribution of borehole data across the study area and, the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.5 Stratified Unit C

Unit description

This unit comprises an unoxidised silty clay, stratified layer that also contains small amounts of coarse sand and gravel. Silt and clay of this unit are dark grey to olive grey (5Y 4/1 to 5Y 5/2). The sand ranges in texture from fine to coarse-grained and is generally well sorted.

Thickness and distribution

SU-C covers the majority of the SWSS although it is not recognised in the very south (south of the Tyner Valley) (Fig 5.15). The units elevation decreases from 745 m along the southern edge of the map area to 467 m in the North Battleford Valley. This sequence is thickest within the main Tyner Valley and its northern tributary, where it can reach a thickness of 38 m (Fig 5.15).

Differentiation from other units

SU-C can only be differentiated from other stratified units where it overlies DU-3 and where it is in turn overlain by DU-4. As a result, in the south of the study area, where DU-4 is absent, SU-C cannot be identified. In order to be consistent when identifying units, the approach taken by previous authors (Christiansen, 1987; Andriashek and Fenton, 1989) within the Canadian Prairies is adopted. Thus it is assumed that all clay, silt, sand and gravel deposits that overlie DU-3 and that are overlain by DU-5/6 are the stratigraphically higher SU-D and E. It is therefore possible that in the south of the study area some of the sediment included in SU-D/E is actually SU-C.

Nature of contact

In most places the lower contact of this unit is sharp. However, within the uplands in the southeast, silt and clay of SU-C rests on the unoxidised surface of DU-3 indicating an unconformable lower contact. In contrast to its lower contact, the upper contact of SU-C is commonly gradational with the base of DU-4 and 5.

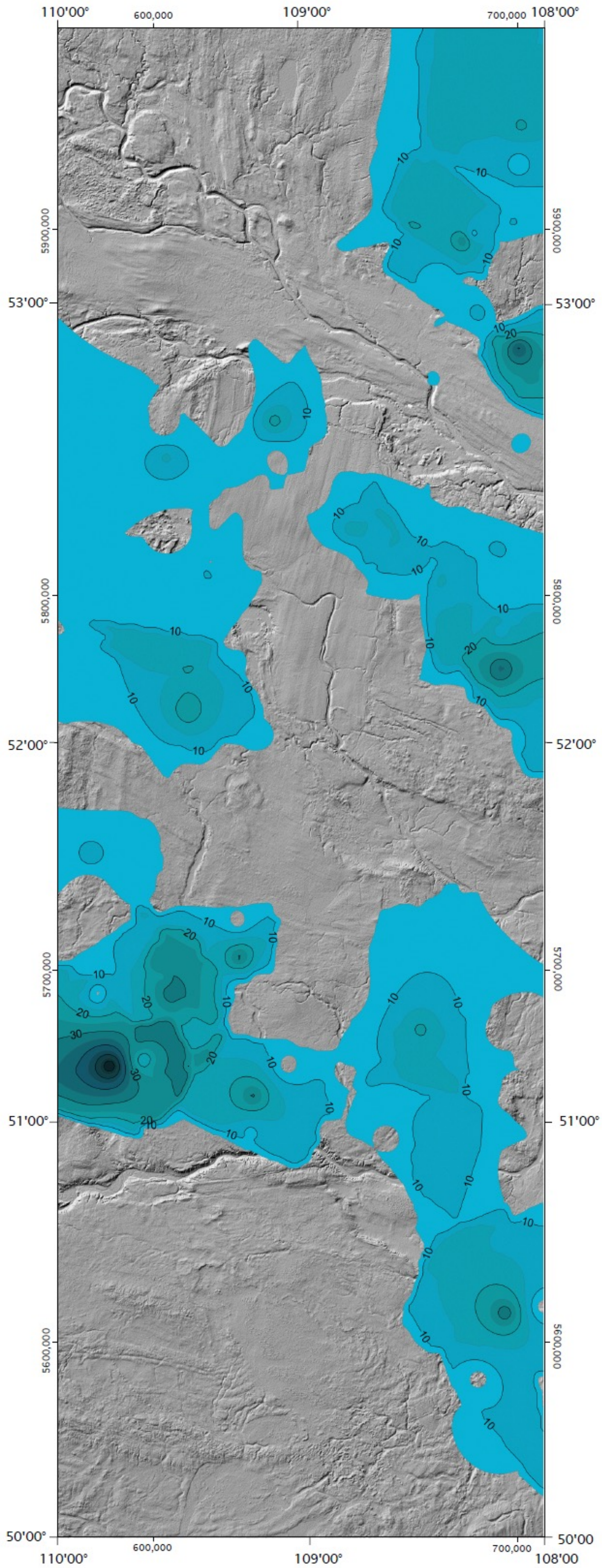


Figure 5.15: *Thickness of silt and clay (small amounts of sand and gravelly sand) comprising SU-C. Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.6 Diamicton Unit 4 and Stratified Unit D

Unit description

DU-4 is composed of dark grey (5Y 4/1 to 5Y 3/1) sandy-clay diamicton. In many boreholes silt and clay are found. In places these silts and clays are embedded in a sandier diamicton thus bearing much similarity to the underlying DU-3. Within the unit, large amounts of calcareous material, particularly dolostone in the sand fraction and dolomite in the silt-clay fraction, are recorded. In a small number of boreholes the top of DU-4 is recognised by the occurrence of a thin layer of sand and gravel (SU-D).

Thickness and distribution

DU-4 extends in a lobate form southwestwards, terminating ~ 110 km south of the Lloydminster Channel and is generally thin within this northeasterly region of the SWSS, particularly within the Turtlelake Upland, where it is commonly between 4-15 m. The unit thickens within two regions. Firstly the thickness of the unit increases within the North Battleford Valley where as much as 30 m are present. Secondly, this unit also thickens towards its southern limit, reaching a total thickness of 46 m (Fig 5.16). It should be noted that the location of this limit coincides with a large area of arcuate overridden ridges, TR-3 (Map Sheet 1).

Differentiation from other units

DU-4 is differentiated from DU-5 based on its substantially more clayey nature and the common appearance of small amounts of silt and clay. Additionally DU-4 is differentiated from the underlying DU-3 due to the considerably more calcareous nature and the coarse sand composition of this unit (Fig 5.17).

Nature of contact

The contact between DU-4 and the underlying silt and clay of SU-C is gradational (see Section 5.2.5). Within multiple boreholes located in the North Battleford Valley SU-C is absent. In these areas DU-4 overlies DU-3, and a sharp contact is recorded (Fig 5.17). A gradational upper contact is recorded in the majority of boreholes between DU-4 and 5. In a small amount of boreholes a thin >13 m layer of stratified sand and gravel (SU-D) separates DU-4 from DU-5, although this unit is very localised.

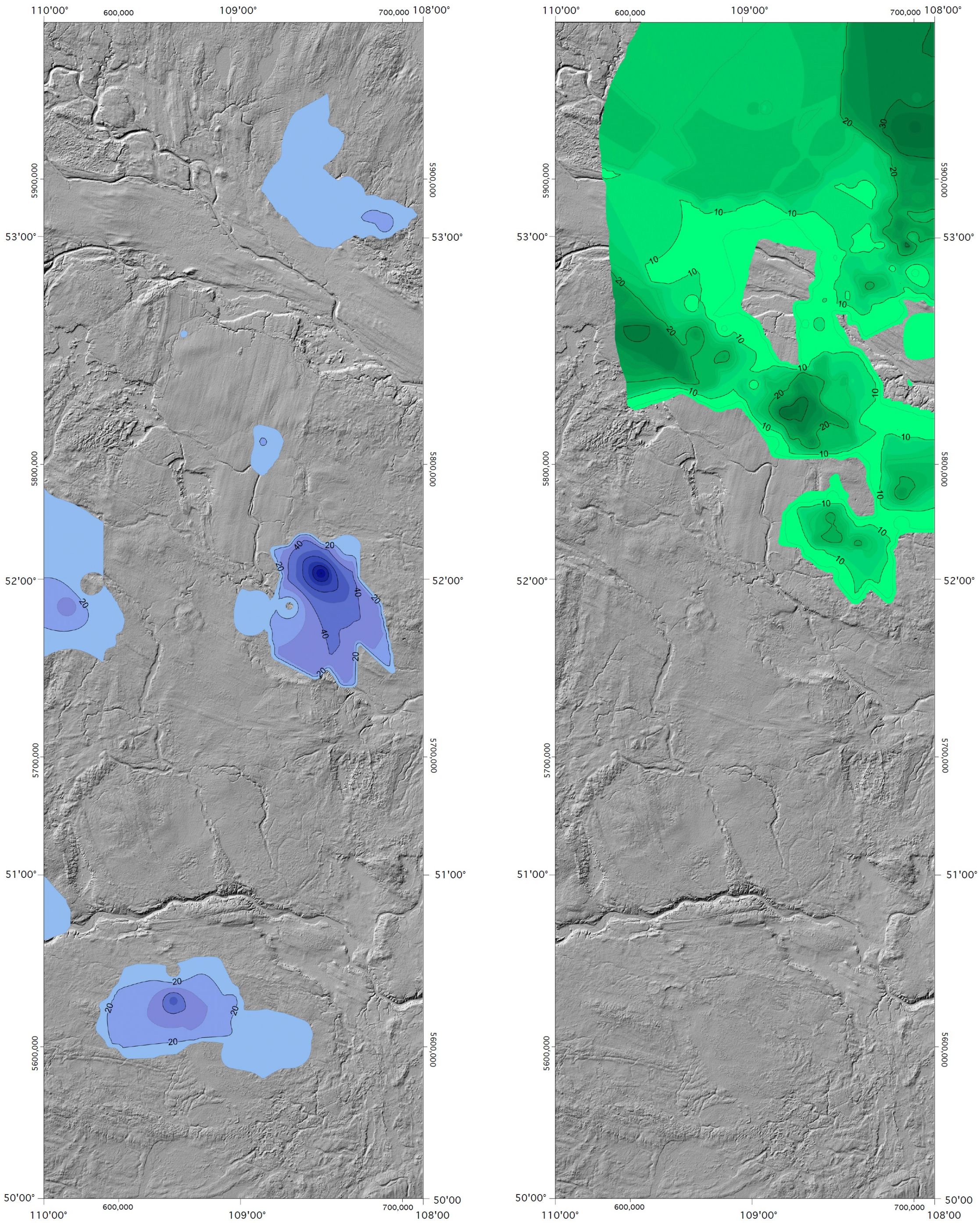


Figure 5.16: *Thickness of sand and gravel comprising SU-D (left) and sandy-clay diamicton comprising DU-4 (right). Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

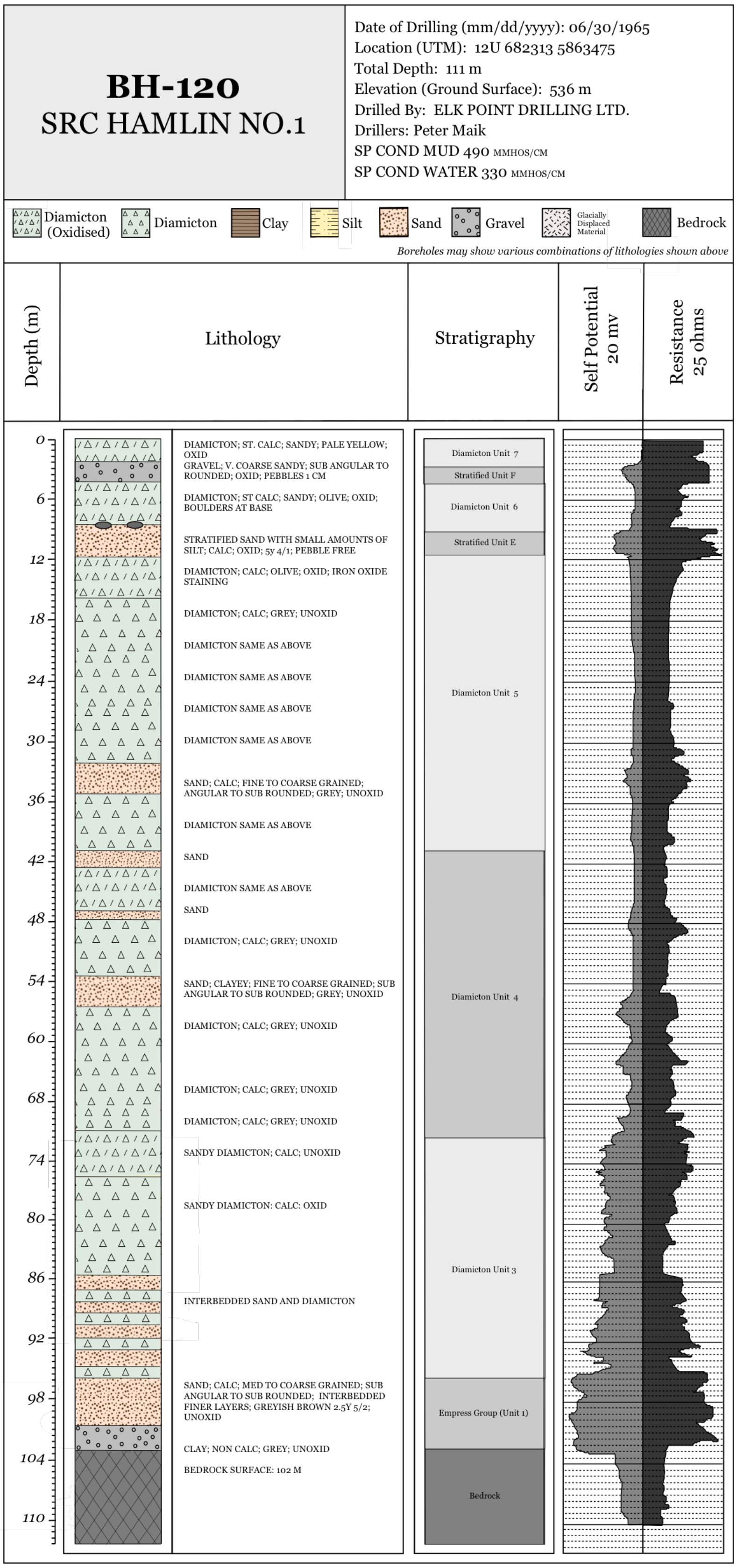


Figure 5.17: *Composite borehole log (Borehole ID-120) from the SWSS. Note the distinct change in electric log response between DU-4 and 3 resulting from a change from clay rich to sand rich diamicton.*

5.2.7 Diamicton Unit 5

Unit description

The unit covers the majority of the SWSS and is composed mainly of sandy diamicton, the top of which is oxidised olive (5Y 4/3) to olive brown (2.5Y 4.5). Below this oxidised zone, the diamicton is dark grey (5Y 4/1 to 5Y 5/1). Like the underlying DU-4 this unit is also rich in calcareous material, particularly dolostone. Iron oxide or manganese oxide staining is recorded. In several boreholes, boulders are recognised at the unconformable contact between DU-5 and the overlying DU-6. It is uncertain at this stage as to which unit such boulder concentrations should be associated.

Thickness and distribution

DU-5 is widespread within the study area. It is considerably thicker than the underlying DU-4. However, like DU-4, it shows a generalised thickening towards the southwest although it is also thick within the Battleford Valley system. The elevation of DU-5 ranges from as high as 799 m within Western Hills Upland to as low as 468 m within the North Battleford Valley. The unit is between 10-20 m thick over much of the northeast and thickens to 20-30 m within the southwest (Fig 5.18).

Differentiation from other units

DU-5 is differentiated from the overlying DU-6 based on its colour, structure, grain size and resistivity. The top of DU-5 is generally more olive brown, in contrast to the dark grey brown of DU-6. In almost all boreholes, DU-5 has a distinctly higher resistivity compared to the base of the overlying DU-6 (Fig 5.17). This probably results from the combined effects of the oxidised zone and the abundant sand at the top of DU-5.

Nature of contact

The contact between the DU-5 and SU-C is discussed in the previous section (Section 5.2.6). In a small number of boreholes in the south of the study area DU-5 directly overlies DU-3. In such cases the contact is sharp and easily recognised on resistivity logs in the majority of places. The upper contact between DU-5 and the clay, silt, sand and gravel of SU-E is generally sharp. Within several boreholes this contact is accompanied by boulders, although it is uncertain as to which unit they belong.

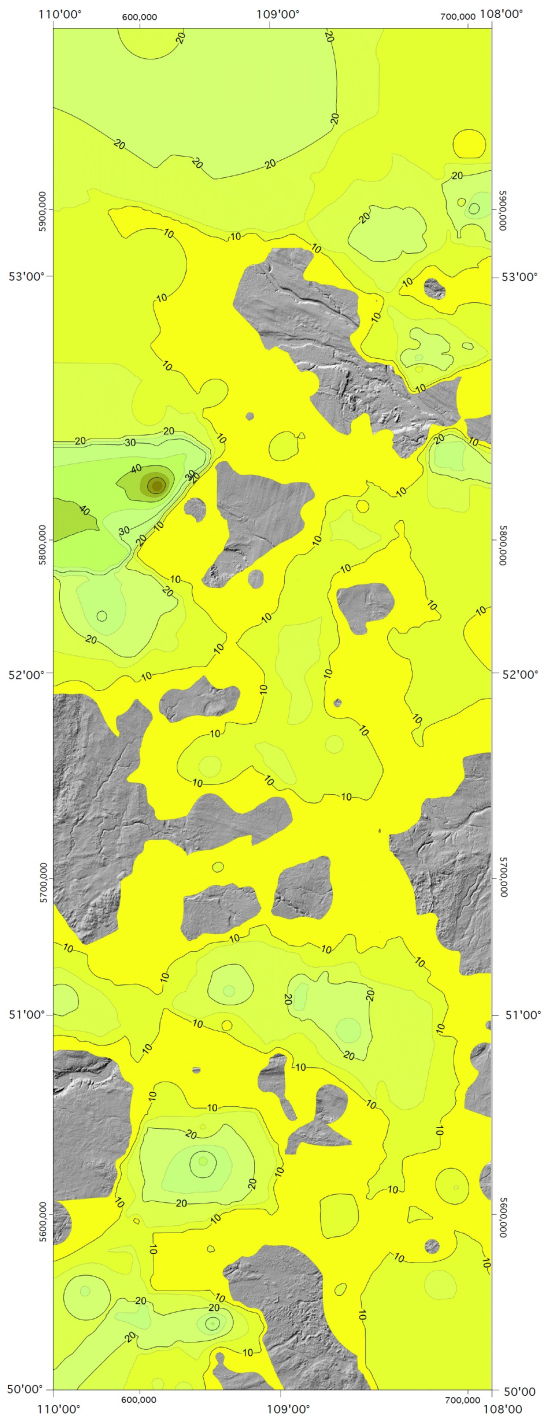


Figure 5.18: *Thickness of sandy diamicton comprising DU-5. Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.8 Stratified Unit E

Unit description

The composition of SU-E is dominated by sand and in two locations (the eastern part of the North Battleford Valley and north of the Western Hills Upland) a lower layer of silt and clay is present (Fig 5.17). Silt and clay that are recorded in this unit are described as unoxidised. In contrast the sand that makes up the majority of the unit is dark grey (5Y 4/1 to 5Y 5/1). It is also generally fine to medium-grained, well sorted and commonly free of clasts. While the composition of sand in this unit was not examined, drillers' logs indicate it is mainly quartz, but may also contain small amounts of igneous, metamorphic and locally derived rock.

Thickness and distribution

Figure 5.19 shows that SU-E forms a discontinuous layer across DU-4 and 5 in lowland regions of the study area, with thick parts of the unit infilling depressions on the surface of DU-4 and 5, especially in the Tyner and Battleford buried valley systems. The upper surface of the unit varies in elevation from 779 m in the Cypress Hills to 494 m in the North Battleford Valley.

Differentiation from other units

Like many of the stratified units in this study, SU-E displays no distinct lithologic or compositional properties that differentiate it from other stratified units. Thus SU-E is differentiated and correlated mainly by stratigraphic position. It should also be noted that sediment in this layer might also be the stratigraphically lower SU-C. As stated in section 5.2.5, in cases where not all diamicton units are present and stratified units are not easily identifiable, in order to be consistent when identifying units, it is assumed that all clay, silt, sand and gravel deposits that overlie DU-3 and that are overlain by DU-5/6 are the stratigraphically higher SU-D and E. It is therefore possible that in the south of the study area some of the sediment included in SU-D/E is actually SU-C.

Nature of contact

In the majority of locations SU-E has a sharp contact with diamicton. However the contact is conformable where oxidised sediment of the SU-E overlies oxidised till of DU-4 or 5. i.e. the eastern part of the North Battleford Valley and north of the Western Hills Upland. The upper contact of this unit can be either sharp or gradational. A sharp contact occurs where unoxidised sediment of DU-6 overlies oxidised sand of SU-E. However, gradual contacts occur where the upper part of SU-E consists of sand interbedded with diamicton that is lithologically similar to the material at the base of the overlying DU-6 (Fig 5.17).

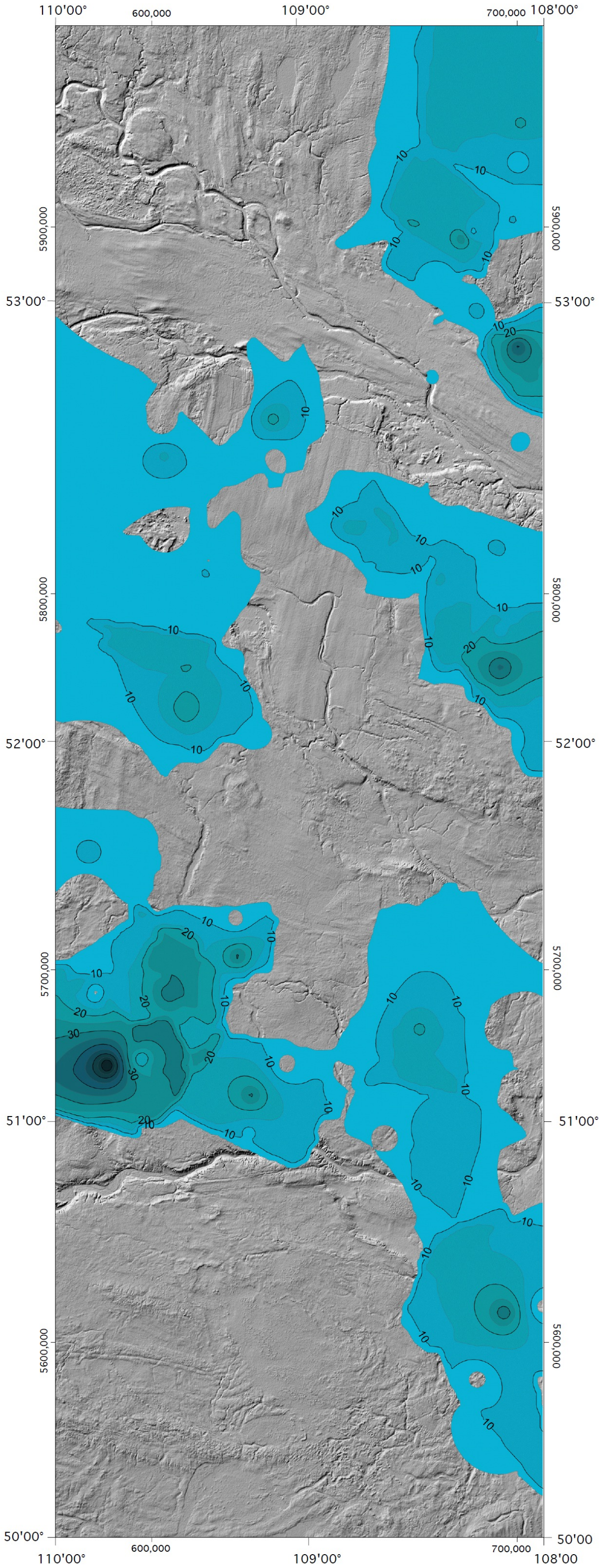


Figure 5.19: *Thickness of stratified sand and silt (with lesser amounts of clay and gravel) comprising SU-E. Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.9 Diamicton Unit 6 and Stratified Unit F

Unit description

DU-6 is described on drillers' logs as 'soft, massive and unstained'. It is mainly composed of sandy diamicton, which is oxidised olive brown (2.5Y 4/4 to 2.5Y 5/6) but becomes very dark grey with depth (5Y 5/1) and is light grey when unoxidised (5Y 7/1) (Fig 5.20). The diamicton is sandy, but contains few clasts (although boulders are recorded at the base and as distinct concentrations within the unit) (Fig 5.20). Glacial streamlined landforms on the surface of this unit have a characteristic north-south orientation. Where this unit is overlain by DU-7, a thin (0.5-4 m) layer of stratified sediment (SU-F) is recorded within numerous boreholes (Fig 5.21).

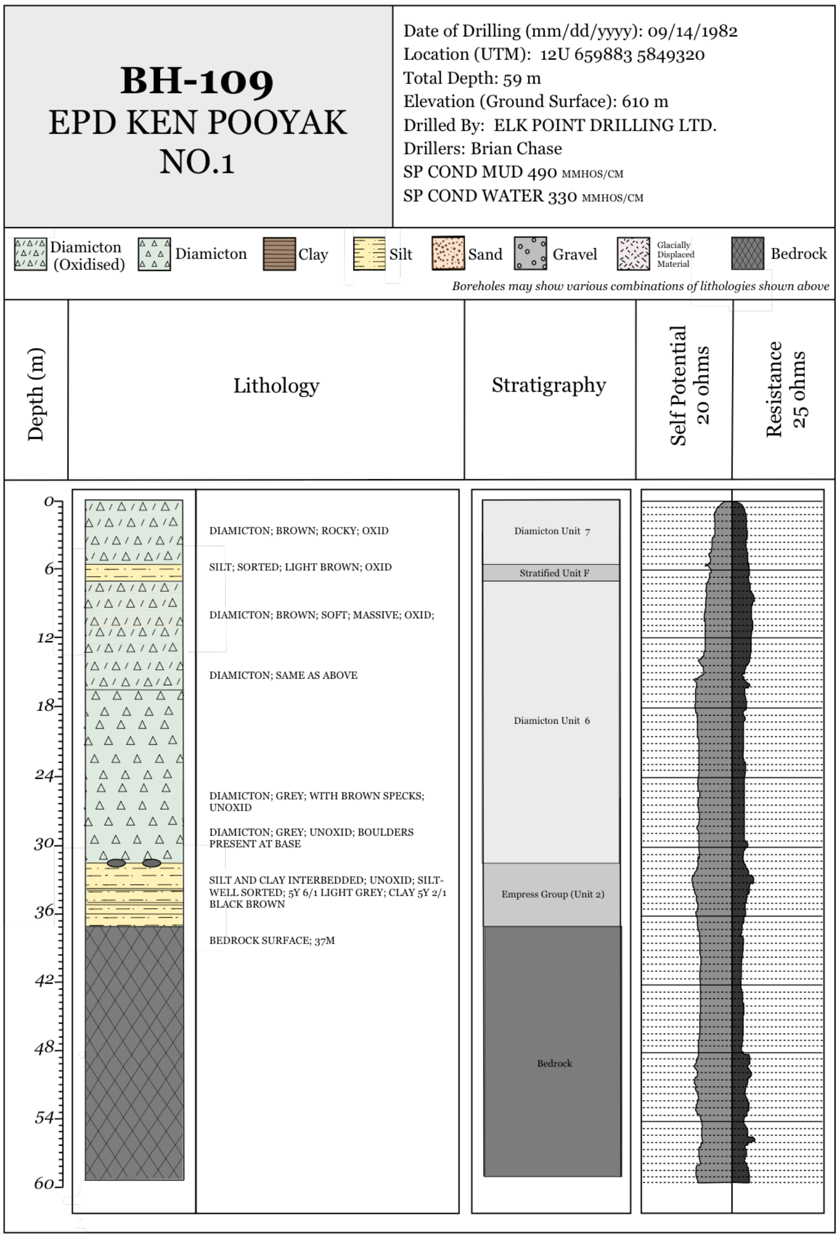


Figure 5.20: *Composite borehole log (Borehole ID-109) from the SWSS. Note the occurrence of boulders recorded at the base of DU-6.*

Thickness and distribution

DU-6 extends across the majority of the study area. Averaging 25 m thick (range 30-3 m) over large parts of the SWSS, it is one of the thickest stratigraphic units (Fig 5.21). The unit thins along the central strip of the study area in correspondence with many areas of streamlined glacial landforms. The unit also displays a generalised thickening towards the south of the study area, reaching thicknesses of 20-30 m (Fig 5.21). Additionally the unit also thickens within the Battleford and Tyner buried valley systems and thins immediately south of the North Battleford Valley.

Differentiation from other units

DU-6 is differentiated from the overlying DU-7 based on: **1.** stratigraphic position; **2.** the considerably more sandy and lighter brown coloured nature of DU-6; and **3.** the fact that glacially streamlined landforms on the surface of DU-7 have a northwest-southeast orientation in contrast to the north-south features on the surface of DU-6. In many places DU-6 was difficult to differentiate from DU-5/4 because these units are sandy and display similar colours. However DU-6 contains few clasts and drillers' logs record a lower carbonate content. It should be noted that in some cores DU-6 shows much similarity to the lower DU-4 and 5. In these cases borehole logs were compared using 3D views in Rockworks to the surrounding boreholes to enable the most likely interpretation, however it is possible that in a small number of cases the diamicton making up DU-6 also contains DU-4 and 5.

Nature of contact

The basal contact of DU-6 is commonly marked by a boulder layer (though this boulder layer is also recognised mid-way through the unit in some cores especially within the north of the SWSS) especially in the south of the SWSS and is distinct and unconformable in most places. The upper contact between this unit and the overlying DU-7 is most easily identified where stratified sediment (SU-F) lies between the diamicton of these two units (Fig 5.20), but in the majority of cores this contact is sharp irrespective of the presence of SU-F.

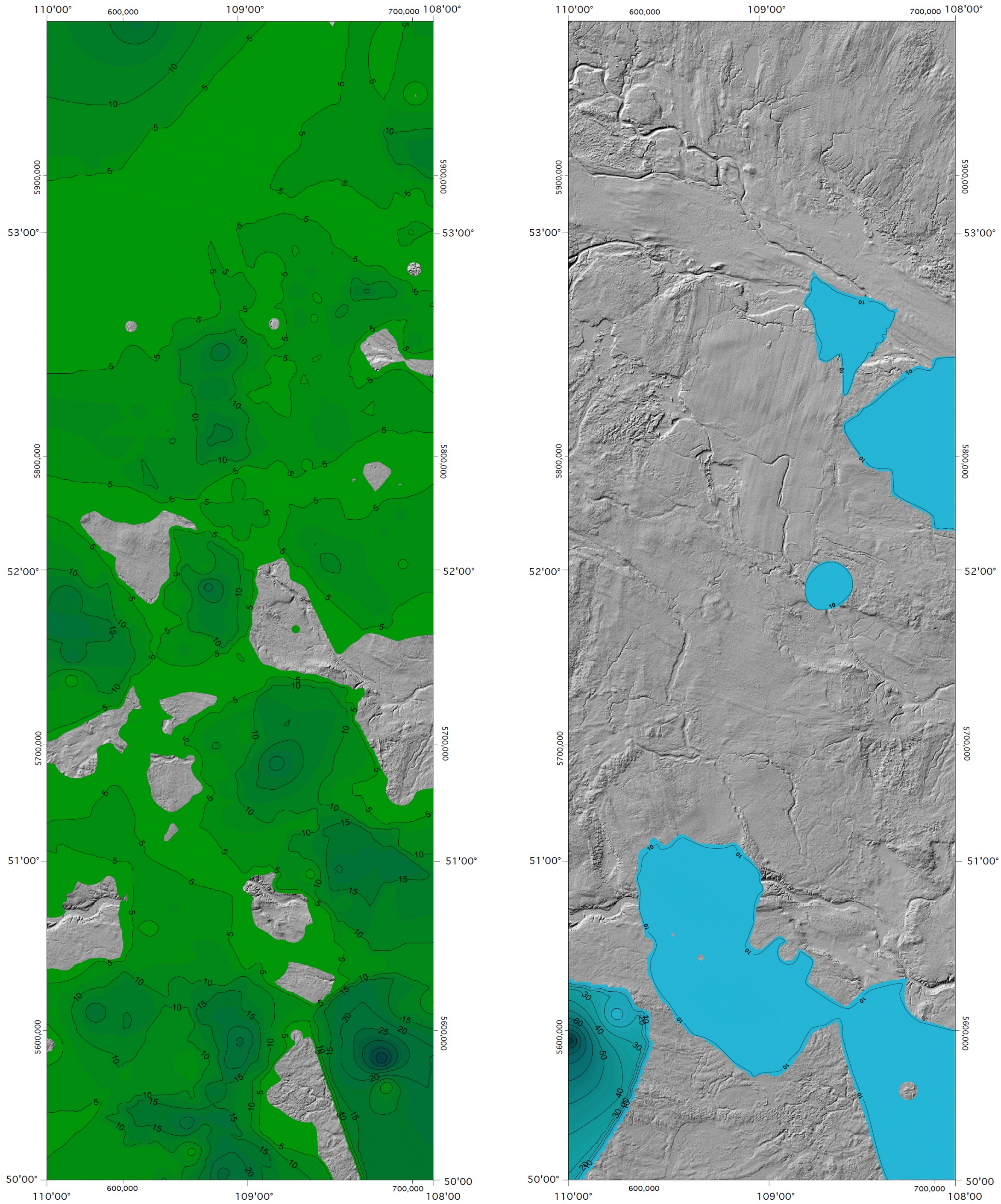


Figure 5.21: *Thickness of sandy diamicton in DU-6 (left) and stratified sediment comprising SU-F (right). Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.10 Diamicton Unit 7

Unit description

This unit is composed of clayey diamicton and in some cores contains glacially displaced older sediment. The unit is oxidised dark grey brown (2.5Y 4/2 to 2.5Y 5/2) in the upper part but becomes very dark grey (2.5Y 3/1) with depth (Fig 5.17). Glacially streamlined landforms on the surface of the unit are well defined and have a northwest-southeast orientation (Map Sheet 1).

Thickness and distribution

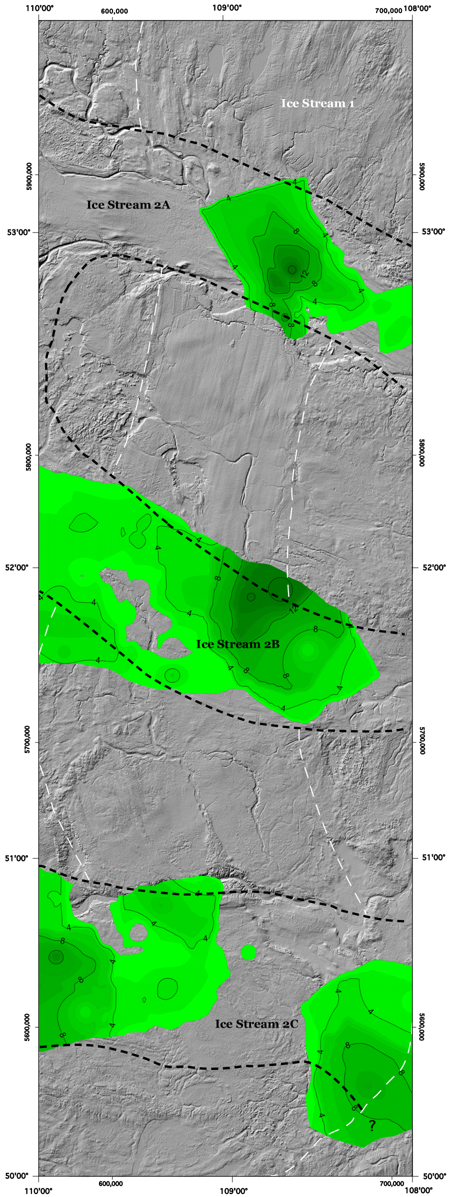
DU-7 is restricted in distribution to three west-east strips that dissect the study area (Fig 5.22). The boundary of this unit is not easily defined mainly due to the lack of sufficient boreholes is the northwest and southeast. The surface topography is thus most useful for establishing the units' boundaries and is defined by the boundary of streamlined landforms and in some areas lateral ridges. The top of the unit ranges in elevation from as high as 782 m within the central corridor to as low as 467 m. DU-7 is thickest within the North Battleford Valley (20 m). The thinnest sediments (<1 m) of the unit are found along the northern margin of the central west-east strip where the unit exhibits a discontinuous layer (Fig 5.22).

Differentiation from other units

As discussed in section 5.2.9, DU-7 is differentiated from the underlying DU-6 based on: **1.** stratigraphic position; **2.** the considerably less sandy and darker brown coloured nature of DU-7; and **3.** the fact that glacially streamlined landforms on the surface of DU-7 have a northwest-southeast orientation in contrast to the north-south features on the surface of DU-6.

Nature of contacts

The contact between DU-7 and 6 is discussed in section 5.2.9. The upper contact between DU-7 and the SSD is conformable and gradational especially where glaciolacustrine deposits are incorporated within till. In such cases the top of the uppermost till lens are selected as the contact.



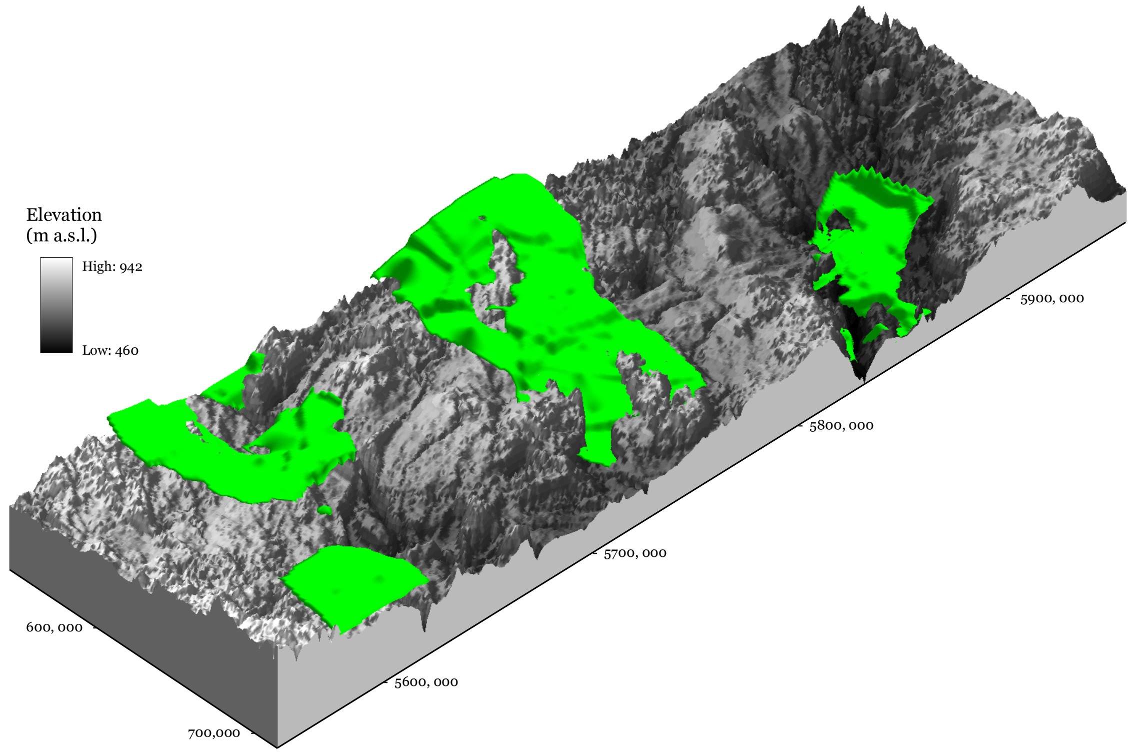


Figure 5.22: *2D and 3D (SRTM draped) representation of the thickness of clayey diamicton comprising DU-7. Due to the irregular distribution of borehole data across the study area and the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.2.11 Surficial Stratified Deposits

Description and distribution of unit

Surficial stratified deposit is the term designated for all sediments situated above the uppermost diamicton and the land-surface. Occurring in the majority of boreholes, this unit is made up of sand, silt and clay. The unit ranges in thickness from <1 m-100 m, however on average it exhibits a thickness of ~2-10 m (Fig 5.23). SSD's commonly become finer grained with depth and display a gradational and conformable contact with DU-6/7 (5.6).

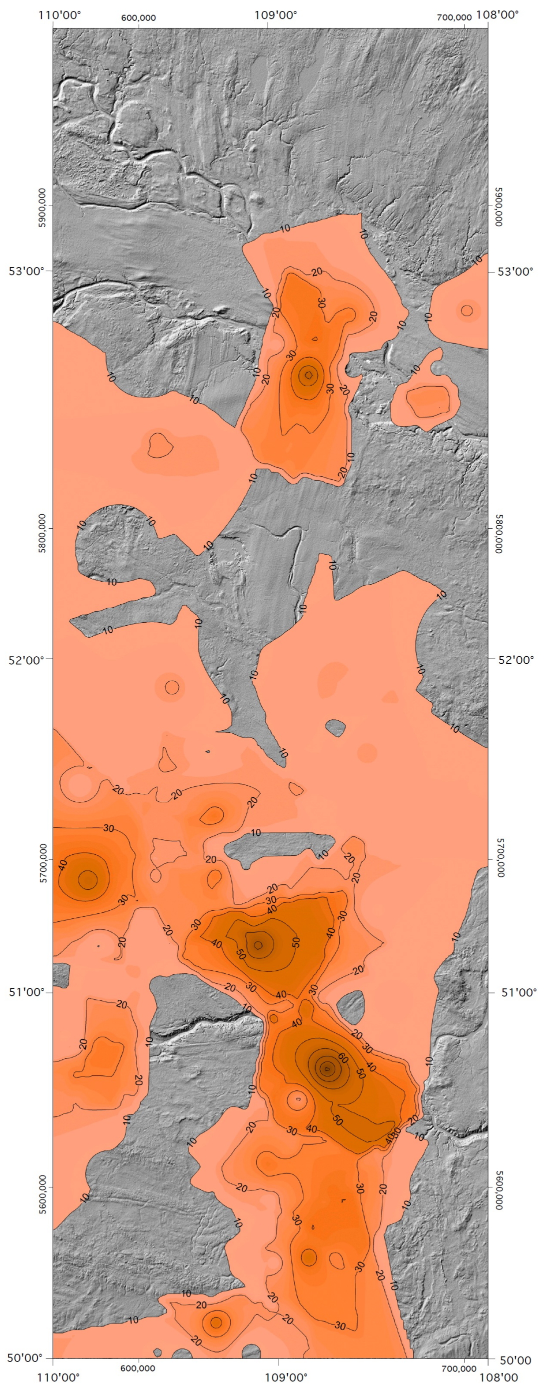


Figure 5.23: *Thickness of silt, sand and clay comprising SSD. Due to the irregular distribution of borehole data across the study area and, the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

5.3 Summary

Large scale geomorphological mapping has allowed the identification of a large corridor of streamlining cross cut by 3 smaller corridors of highly attenuated bedforms. Within each of these corridors a variety of transverse ridges, sinuous ridges and channels are identified. Stratigraphic modelling reveals the occurrence of 7 diamicton units separated and underlain by 10 stratified units. Within the following chapter the components of this landscape are analysed, before their significance is discussed in Chapter 7.

6. Interpretation of Landforms and Sediments

The subglacial landscape and the surface geomorphology of the SWSS are consistent with a sediment-landform model that involves the evolution of multiple ice stream systems. These assemblages are now further analysed in the following sub-sections to provide insight into palaeo-ice stream characteristics and to reconstruct the evolution of these systems in the context of the Late Wisconsinan deglaciation.

6.1 Glacial geomorphology

6.1.1 Megageomorphology and lineations (glacially streamlined bedforms)

In agreement with Ross et al. (2009) and Ó Cofaigh et al. (2010) the smoothed cross cutting corridors 1 and 2A, B and C (Map Sheet 1) are interpreted as zones of former subglacial streamlining within the southwest margin of the LIS, produced in a region where ice flowed at higher velocities than the surrounding ice (Swinthinbank, 1954; Bentley, 1987). Thus this study hereafter refers to corridors 1 and 2A, B and C as palaeo-ice stream tracks 1 and 2A, B and C (see Fig 6.1 for locations)

These corridors are surrounded and delineated by a change from the ‘smoothness’ of the streamlining (Ó Cofaigh et al., 2010) to the hummocky terrain (see Section 5.1.3) associated with slow moving, cold based ice and stagnation (Dyke and Morris, 1988, Stokes and Clark, 2002b, Evans et al., 2008; Ó Cofaigh et al., 2010). Comparison of glacial geomorphology mapped in Map Sheet 1 against Stokes and Clark’s (1991, 2001) geomorphological criteria for ice streaming shows a very strong correlation (Table 6.1) and the subglacial corridors probably represent the glacial bed as the product of a single flow event, thereby representative of a ‘rubber stamp’ imprint of palaeo-ice stream activity (Clark and Stokes, 2003). Moreover the MSGLs and smoothed topography directly compare to previously identified palaeo-ice streams in the terrestrial landform record (Patterson 1997, 1998; Stokes and Clark, 1999, 2001; Clark and Stokes, 2003; Jennings, 2006) and to the forelands of contemporary ice streams on the Antarctic Shelf (Shipp et al., 1999; Canals et al., 2000; Wellner et al., 2001; Ó Cofaigh et al., 2002).

Table 6.1: *Ice streams in the SWSS compared to ice stream geomorphological criteria proposed by Stokes and Clark (1999, 2001).*

Ice stream geomorphological criteria (Stokes and Clark, 1991, 2001)	Ice Stream 1	Ice Stream 2A	Ice Stream 2B	Ice Stream 2C
1. Characteristic shape and dimensions	YES	YES	YES	YES
2. Convergent flow patterns	YES	YES	YES	Unknown
3. Highly attenuated bedforms	YES	YES	YES	YES
4. Boothia-type erratic dispersal trains	Unknown	Unknown	Unknown	Unknown
5. Abrupt lateral margins	YES	YES	YES	YES
6. Lateral shear margins	YES	YES	YES	YES
7. Evidence of pervasively deformed till	YES	YES	YES	YES
8. Submarine till delta or sediment fan	N/A	N/A	N/A	N/A

Ice Stream 1 is the largest palaeo-ice stream mapped in the study area with a length of >600km. When compared to bedrock and surface topography maps (Fig 6.3) it is apparent that its flow path developed against the regional slope and was unconstrained by topography (Tyner and Battleford buried valley systems) that it cross cuts. Thus it is considered to be a pure ice stream (Stokes and Clark, 1999).

Within the study area this ice stream exhibits a convergent flow signature, displaying a ~105 km wide zone followed by a narrow trunk zone (~58 km wide) and a progressive widening to the terminus (~170 km wide). The zone of flow convergence north of the flow path of Ice Stream 2B (between 53' 00° and 52' 40° N) is marked by MSGLs that are highly attenuated and range from 11-60 km long. The lateral boundaries of this heavily streamlined zone are demarcated by prominent lateral moraines (Map Sheet 1). Based on comparison with the spatial signature of modern ice streams, in which an increase in flow convergence is typically associated with an increase in flow rate, it is inferred that this region of the ice stream was characterised by high velocities (Rignot, 2006; Joughin et al., 2001). This inference is also supported by the highly attenuated appearance of bedforms in this zone of the SWSS, which have also been linked to fast flow (Dyke and Morris, 1988; Clark, 1993; Wellner et al., 2001; Ó Cofaigh et al. 2002; Briner, 2007).

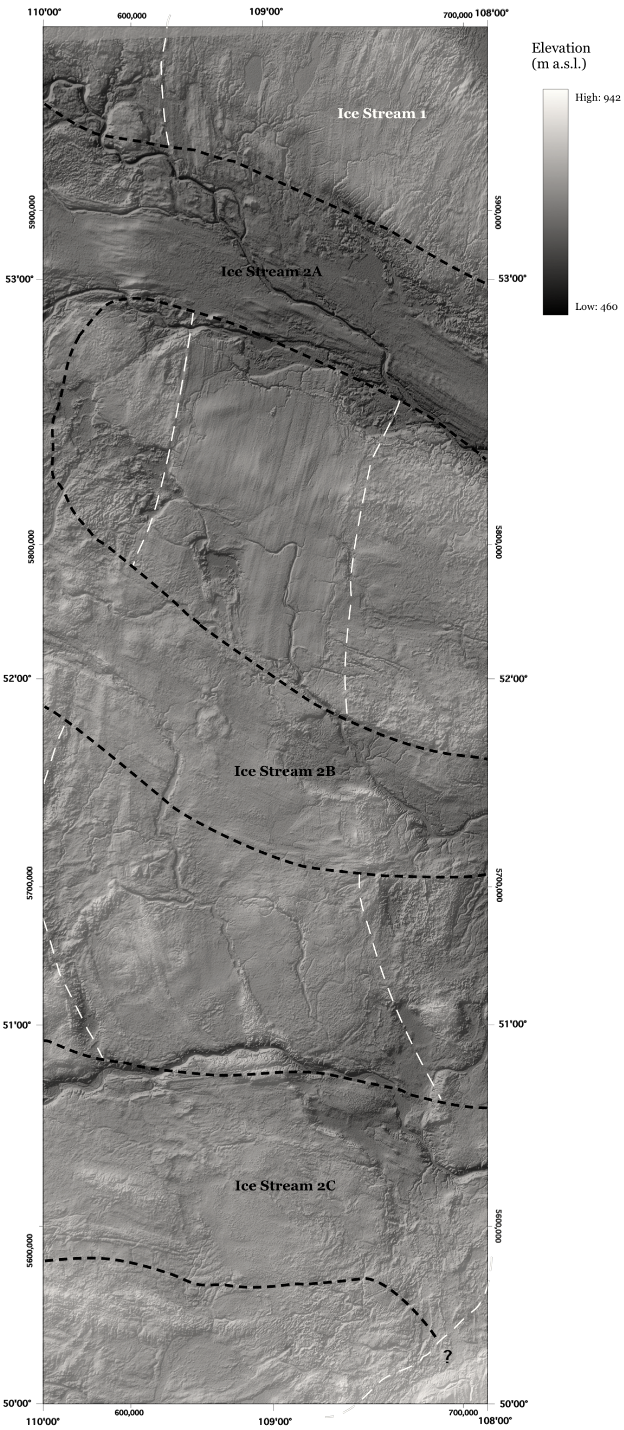


Figure 6.1: *SRTM-90 DEM showing the locations of smoothed cross cutting corridors. White dashed line demarcates Ice Stream 1; white dashed line demarcate flow path of Ice Stream 2A, B and C. Note the topographic confinement of Ice Streams 2A, B and C.*

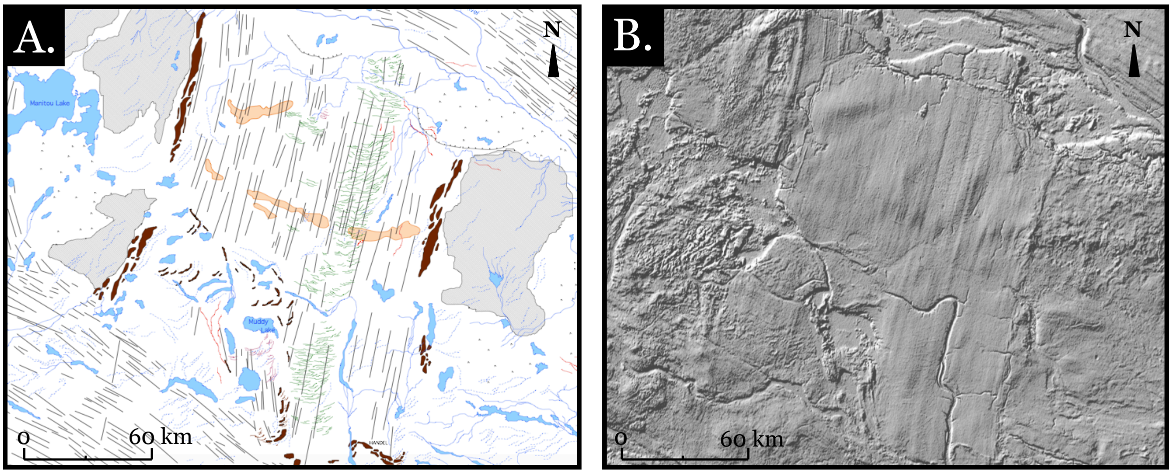


Figure 6.2: *Extract of geomorphic mapping (see Map Sheet 1 for full map) and SRTM imagery of the change in 'smoothness' from a smoothed corridor of lineations representative of fast ice flow to a region of hummocky terrain associated with slow moving or stagnant ice.*

In three zones of the study area the north-south flow path of Ice Stream 1 is crosscut at 90° by the topographically confined Ice Stream 2. The reconstruction of this ice stream is made based on the recognition of three major flow-sets, 2A, B and C. Ice Stream 2A records southeastward streaming confined within the buried North Battleford Valley as defined by a concentration of 8-17 km long MSGLs aligned down the valley's long axis. Flutings within Corridor 2A have been previously discussed by Gravenor and Meneley (1958) and again by Grant (1997) who referred to the area as the 'North Battleford Fluting Field'. However the density of MSGLs they identify in both of these studies is significantly less than identified here (Fig 6.4). This discrepancy is attributed to the lack of high resolution imagery and local scale of both of these studies. The flow path of this ice stream was up to 40 km wide and at least 180 km long, although this is a minimum estimate based on the distribution of MSGLs along the valley, it is possible that Ice Stream 2 extended much farther to the northwest. Evidence for this is proposed by Andriashek and Fenton (1989) and Evans et al. (1999) who map lineations (Lac La Biche Lobe) in the bordering Sand River area 73L of Alberta (54 ° to 55 °, 110 ° to 112 °), showing a correlation with Ice Stream 2 (Fig 6.5). Ice Stream 2B also records southeastward flow. This area of streaming is also topographically confined (Map Sheet 1: cross section A-A') shown by the 50 km wide zone of MSGLs bordered by fragmented lateral moraines. The MSGLs in this corridor are superimposed by several moraine lobes, and ice thrust ridge concentrations (mentioned but not discussed by Evans et al. (2016)).

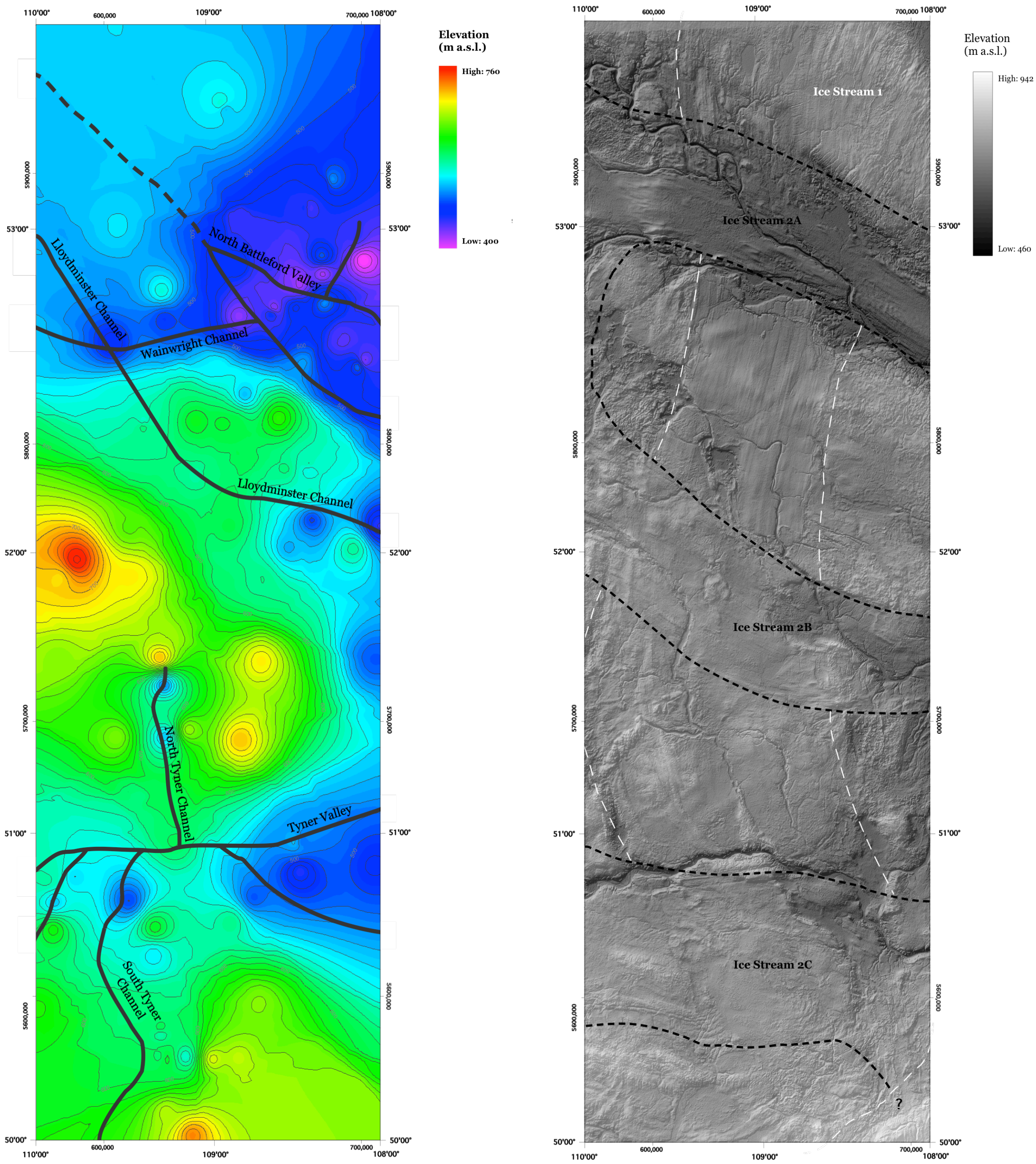


Figure 6.3: *Bedrock topography of the SWSS, and SRTM imagery demarcating ice stream tracks. Comparison of Ice Stream 1 to the bedrock topography map shows the lack of topographic constraint associated with this ice stream.*

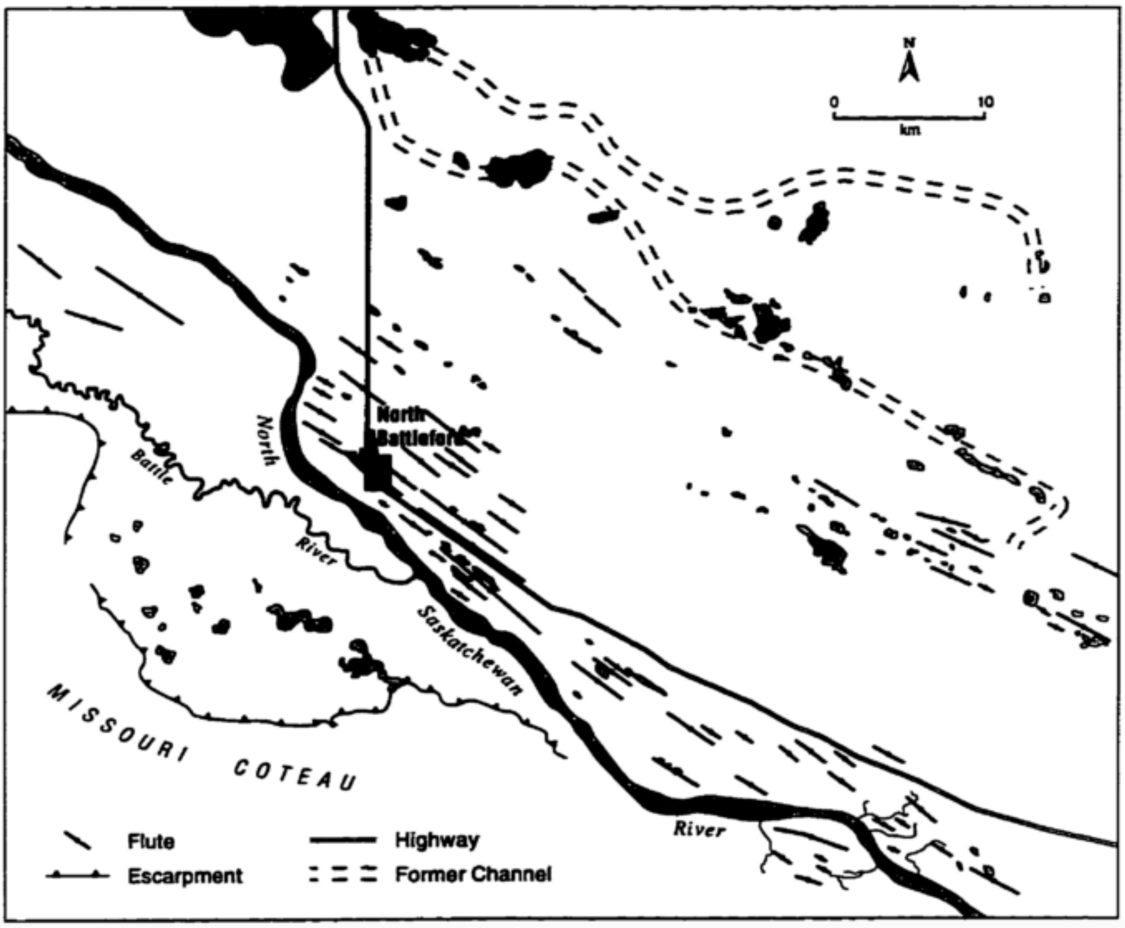


Figure 6.4: *Geomorphological map of the 'North Battleford Fluting Field' and associated features. Taken from: Grant (1997).*

Ice Stream 1 is further crosscut towards its southern end by Ice Stream 2C (Fig 6.6). Previous research which solely used SRTM imagery (Ross et al., 2009; Ó Cofaigh et al., 2010) has not identified this 50 km wide area of MSGLs, although earlier mapping by Campbell (1986a, b, 1987a, b) does identify linear drumlinoid features within the same region. Mapping in this study shows this ice stream displays fewer MSGLs, especially in the centre of its flow path when compared to Ice Streams 2A and B. In this central region a large area of sand dunes now exists. Thus it is possible that MSGLs were formed, but have been superimposed and/or altered by sand deposition.

6.1.1.2 An alternative hypothesis for the formation of lineations

As discussed in the literature review section of this thesis an alternative hypothesis for the formation of lineations has been previously proposed within the Canadian Prairies. This mechanism states that lineations are the product of subglacial megaflood erosion of basal sediment and the underside of the LIS (Shaw, 1983; 1989, 1994; Shaw and Sharpe, 1987). To date, the only research related to this hypothesis within the SWSS has been within a small >70 km region of the North Battleford Valley. On the basis of form analogy Grant (1996) proposed that features within the 'North Battleford Fluting Field' are erosional forms and the most likely agent of erosion was a subglacial megaflood. This hypothesis is rejected for the formation of lineations within the North Battleford Valley and more generally within the whole of the study area based upon two lines of evidence: **1.** the corresponding narrow distribution, fluted surface morphology and stratigraphic position of DU-7; **2.** the presence, pattern and distribution of CSRs which suggests deposition by a narrow stagnating ice mass (see Section 6.1.2.2).

Figure 6.5: *Mapping of lineations in the Sand River area 73L of Alberta redrawn from Andriashek and Fenton (1989). Note the clear extension of lineation mapping in the Sand River area to the corridor of lineations that demarcate Ice Stream 2A in the SWSS. Major ice stream tracks of the HPIS and CAIS (Evans et al., 2008; Evans et al., 2014) and Manitoba Ice Stream system (Patterson, 1997; Jennings, 2006) have been outlined. See section 7.1.3 for further discussion of this regional ice stream network.*

6.1.2 Transverse Ridges (Types 1-4)

6.1.2.1 Transverse Ridge: Type 1 (moraine ridges)

This study reaffirms previous interpretations (Campbell 1986a, b, 1987a, b; Ross et al., 2009; Ó Cofaigh et al., 2010) of multiple, densely spaced arcuate ridges of only a few meters relief occurring over large areas in the southernmost part of Ice Stream 1. Ridges form broad arc shapes, typical of modern push moraines (Price, 1970; Mathews et al., 1979; Evans and Twigg, 2002; Evans, 2003). These moraines define the lobate pattern of the terminus zone of Ice Stream 1. A small group of transverse moraine ridges, referred to as the Handel Moraine by Evans et al. (2016), are also mapped (see Map Sheet 1 for location). These features occur in close proximity with CSRs and likely formed during a late stage of streaming activity in association with the formation of CSRs. Other ice marginal features are present throughout the streamlined corridors including several well preserved lobate moraines north of Ice Stream 2B. These moraines were likely formed by a small, thin ice lobe, an assertion made based on the fact that these moraines follow the contours of the slope in this region. This indicates that the landform record within the SWSS not only consists of elongated ridges that can be traced over long distances, but also that this subglacial landscape is overprinted locally by younger ice recessional or ice readvance features.

Smaller minor TR-1 also occur within Ice Streams 2A, B and C. These moraines collectively form small lobate arcs in several places. However due to their occurrence close to multiple thrust block moraines and their orientation transverse to lineations in these corridors, they are interpreted as (in combination with other characteristic landforms) indicative of a surge signature produced by ice flowing into Saskatchewan from the northwest (Fig 6.7).

Along Ice Stream 2B and along the trunk zone of Ice Stream 1 the lateral margins are marked sharply by lateral moraines that extend discontinuously for distances of >60 km (Ice Stream 1) and >30 km (Ice Stream 2B). In profile the moraines usually comprise a single asymmetric ridge a few kilometres wide and 20-30 m in height above the adjacent ice stream bed. These features are analogous to shear margin moraines formed in the shear zone between the rapidly moving ice stream and adjacent slower moving or stagnant ice (Raymond et al., 2001; Stokes and Clark, 2002b; Hindmarsh and Stokes, 2008).

6.1.2.2 Transverse Ridge: Type 2 (geometric ridge networks)

Geomorphological mapping identified two types of TR-2. A large assemblage of TR-2 are identified in the bed of Ice Stream 1 and a much smaller concentration of small scale TR-2 are also identified on the bed of Ice Streams 2A and B. The large assemblage in Ice Stream 1 overprints the MSGLs in the trunk zone (Map Sheet 1) and consists of >70 ridges with a prominent WNW-ESE alignment. These geometric ridge networks were first identified by Campbell (1987a, b) on surficial geology maps of North Battleford (73C) and St Walburg (73F) and were variously classified as 'crevasse fillings' and 'minor ridged moraines'. More recently these features have also been discussed by Evans et al. (2016), who interpreted them as crevasse-squeeze ridges (CSRs). Based on the diamictic composition of these ridges, which indicates they were created by subglacial till injection into basal and/or full depth crevasses, a CSR origin is supported here. It is clear based on the preservation of landforms that the development of these ridges took place during internal flow unit reorganisation, immediately prior to ice stream shutdown. The prominent WNW-ESE alignment (with a subordinate WSW-ENE alignment) likely reflects fracture development transverse to former ice flow, which would have been predominantly NNE-SSW based upon MSGL alignment. Due to the narrow corridor and non-arcuate appearance of these features, a traditional interpretation of CSR production related to fracturing within glacier surge lobes (see below) is not valid. Thus Evans et al. (2016) proposed a mechanism whereby CSR corridors relate to lateral shear zones formed between flow units within ice streams not characterised by surging.

The second sub-type of TR-2 are also interpreted as CSRs and are located along the trunk of Ice Streams 2A and B. These CSRs are orientated NNE-SSW and are interpreted to be representative of fracturing and injection of till into crevassed ice formed by surging activity of Ice streams 2A and B (Sharp, 1985; Evans and Rea, 1999; 2003; Evans et al., 2007). The juxtaposition of CSRs with lineations, several hill-hole pairs and thrust block moraines are also interpreted as evidence for ice stream surging. The suite of landforms is thus directly compatible with the landforms of contemporary surging forelands and therefore likely represents a surging landsystem (Evans and Rea, 1999; 2003). The association between ice streaming and surging will be considered in the following chapter.

6.1.2.3 Transverse Ridge: Type 3 (ice thrust ridges)

TR-3 ridges are interpreted as thrust block moraines or ice thrust ridges and hill-hole pairs, an origin that is consistent with their multiple crests and occurrence in close proximity to small (often water filled) depressions (Kupsch, 1962). Where these features are of a sharp crested nature they are suggested to have been formed during ice recession. In contrast, where they appear as streamlined and/or smoothed they are interpreted as features originally constructed during initial ice advance and then overridden.

Data on the internal composition of these thrust features is limited due to the small number of boreholes situated within such features. Thus data only allows description of the types of features that may be present in thrust moraines but does not permit the construction of a clear picture of the origin of sediment and/or bedrock in a particular feature. However these ice thrust ridges comprise a hill of glaciotectonically displaced sediment and/or bedrock with a conspicuous up-ice depression (typically filled with a lake) from where displaced material was derived. The composition of thrust sediment is highly variable. At a given location this sediment may consist of two or more layers of diamicton, sand, clay, sandstone, ironstone or mudstone. The proportion of bedrock in the thrust sediment is controlled mainly by the thickness of the pre-existing Pleistocene sediment. Sediment is generally >50 m thick in the SWSS, and as a result many of the large thrust hills are most likely composed entirely of glacial sediment; a prominent example of such a hill-hole pair occurs immediately north of Ice Stream 2A on the eastern side of Ice Stream 1 (see Map Sheet 1 for location).

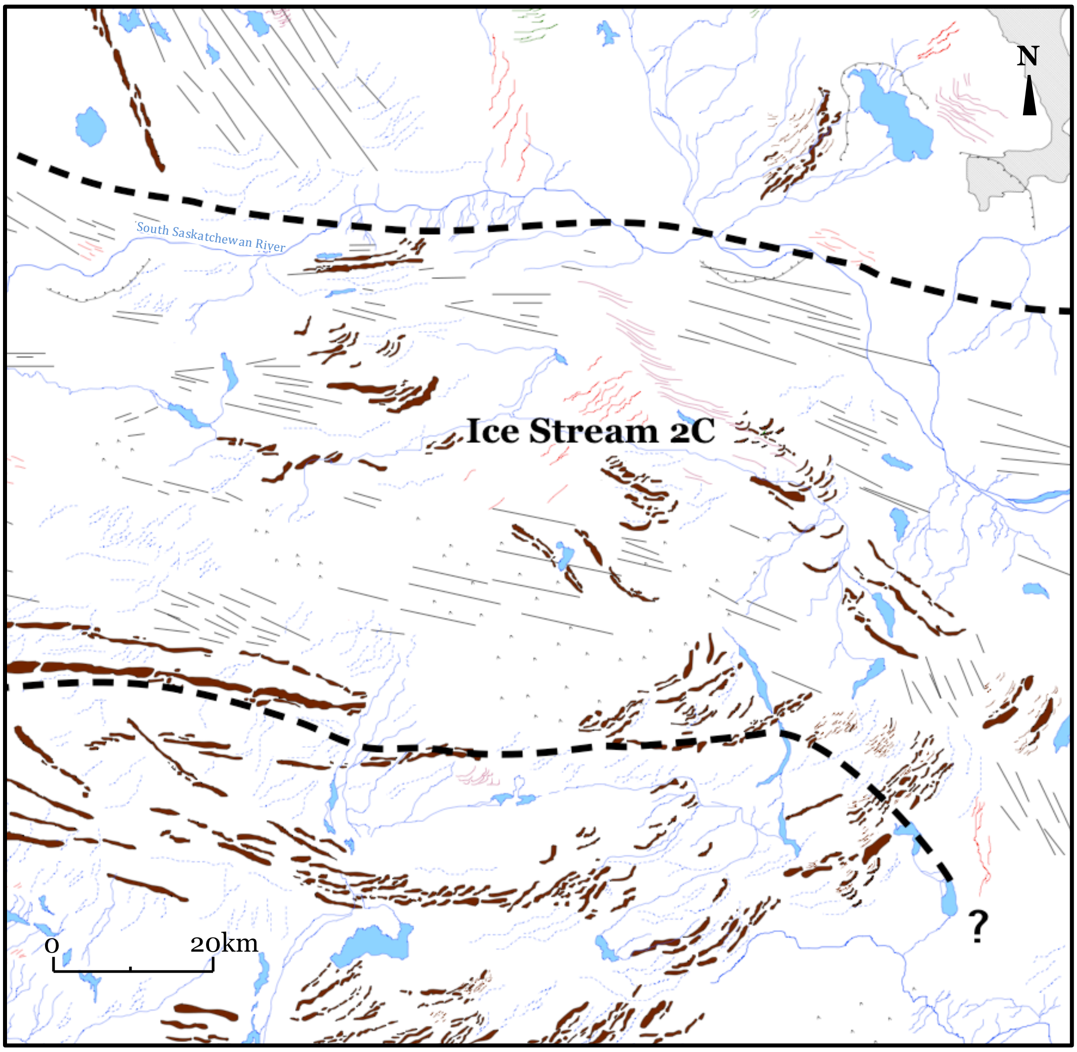


Figure 6.6: *Extract of geomorphic mapping (see Map Sheet 1 for full map) showing the location and landforms demarcating Ice Stream 2C.*

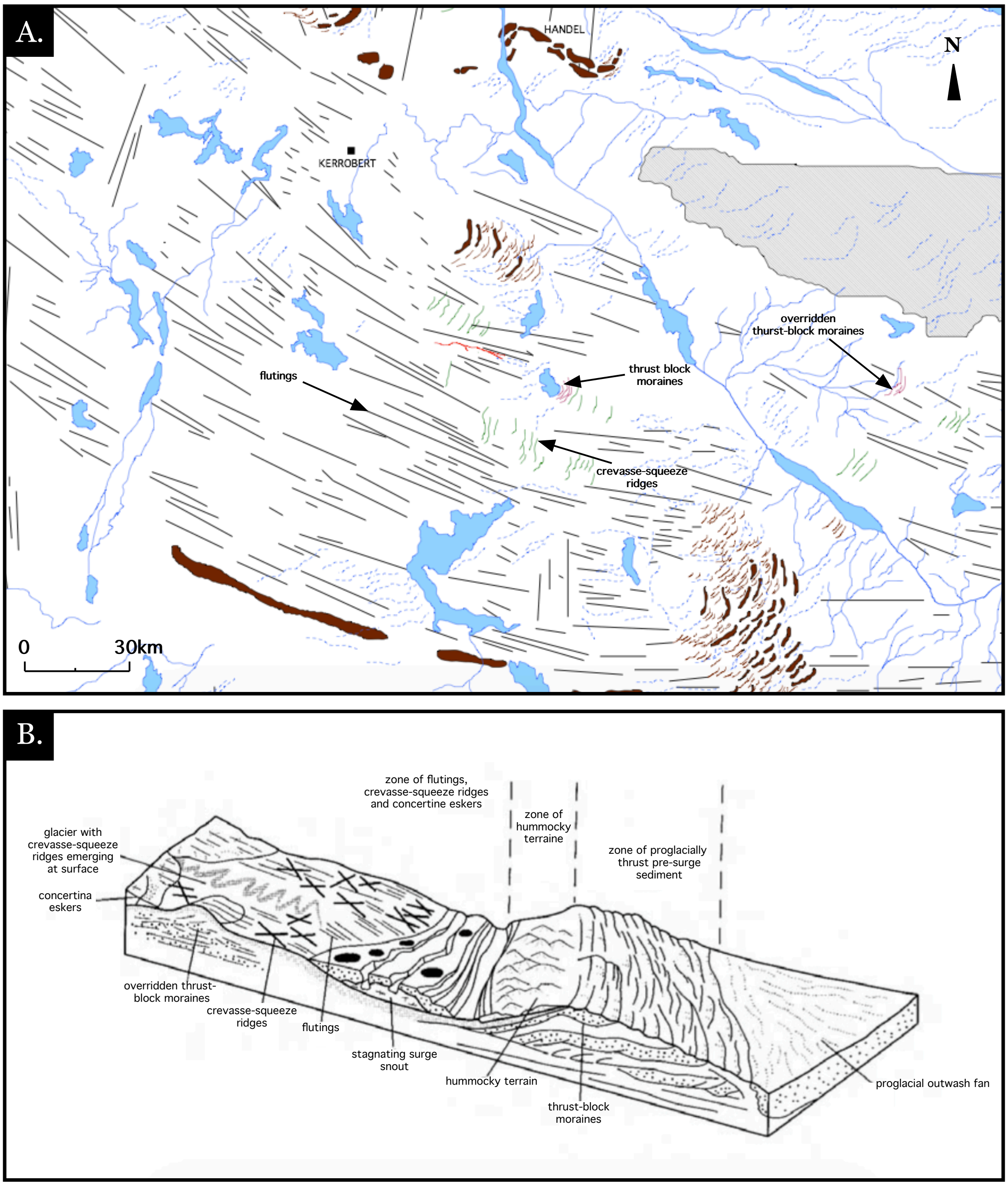


Figure 6.7: *Geomorphic surge signature of Ice Stream 2B.* ***A.*** *Extract of geomorphic mapping (see Map Sheet 1 for full map) showing the elements of a surging glacier margin.* ***B.*** *A landsystem model for surging glacier margins (Taken from: Evans et al., 2003).*

6.1.2.4 Transverse Ridge: Type 4 (overridden ridge complexes)

Within the northern end of the study area several overridden transverse ridge complexes exist. These features form a broadly arcuate lobate shape that correlates to the terminus and thickening of DU-4 (see Section 6.2.6). Thus based on the thickening of this stratigraphic unit, these features are interpreted as ice marginal moraines constructed in association with subglacial till thickening during ice sheet advance.

6.1.3 Inter-corridor and hummocky terrain

Hummocky terrain is the most common landform within marginal zones outside of the flow paths of Ice Streams 1, 2A, B and C. Based on the resolution of SRTM and Landsat ETM+ imagery, these landforms can only be mapped at small scales and thus appear to be chaotic and demonstrate little to no linearity. The abrupt transition from smoothed topography to hummocky terrain along the margins of Ice Stream 1 is interpreted as a change in subglacial regime and hence used to demarcate the flow path of the ice stream (Dyke and Morris, 1988; Patterson, 1998; Evans et al., 2008; Ó Cofaigh et al., 2010). Furthermore previous work in Saskatchewan (Klassen, 1989; Kulig 1996; Cummings et al., 2012) and eastern Alberta (Gravenor and Kupsch, 1959; Stalker, 1960; Bik, 1969) identified a significant proportion of these features are composed of till. A supraglacial origin for these hummocks can be supported by a simple form analogy (Clayton, 1967; Boulton, 1972; Patterson, 1997; 1998; Jennings, 2006), however it should be noted Eyles et al. (1999) have proposed a subglacial origin by elaborating on Stalkers (1960) concept of ice pressing. In this region some subglacial pressing against ice stream margins due to heavily saturated sediments would have been probable (Klassen, 1989). It is therefore likely that hummocky terrain in the SWSS is a polygenetic landform of both supraglacial and subglacial origin.

6.1.4 Sinuous ridges (eskers)

Individual eskers and esker networks have been identified by Campbell (1986a, b, 1987a, b) and by Ó Cofaigh et al. (2010). As previously mentioned, only the largest esker forms were identifiable. North-south orientated, relatively linear, straight limbed eskers occur on the bed of Ice Stream 1. These features are interpreted as being associated with the drainage of this ice stream, with the Glidden Esker Network along Ice Stream 1 documenting that subglacial water was concentrated at this point (Evans et al., 2008). Similarly, the concentration of eskers in the Great Sand Hills Lowland also suggests that subglacial water was concentrated in this area (Map Sheet 1). However due to the orientation of these features being NNE-SSW it is proposed that they formed in association with NNW-SSE flowing Ice Streams 2A, B and C. Where eskers cross cut MSGLs (Map Sheet 1) they are interpreted to have formed after lineation production most likely during ice marginal recession (Stokes and Clark, 2003b; Stokes et al., 2008).

6.1.5 Abandoned channels (ice-marginal meltwater channels and spillways)

Meltwater channels are extensive across the study area, forming continuous chains of channels and water filled depressions, often in close association with esker ridges. Within the flow path of Ice Stream 1 channels are often north-south orientated, such as Tramping Lake (see Map Sheet 1 for location) indicating that they formed subglacially in association with Ice Stream 1. In contrast to these features, meltwater within Ice Streams 2A, B and C are often more continuous and occur in association with large hill-hole pairs. In addition to channels associated with ice stream activity in Ice Streams 1, 2A, B and C, it is also worth mentioning the channel networks within hummocky terrain. Their overall shape resembles that of fluvial drainage networks rather than those created under subglacial hydraulic gradients. The corollary is that they formed as ice stagnated and down wasted in this area of hummocky terrain.

6.2. Stratigraphic interpretations

<u>6.2.1. Origin of the Empress Group</u>

Situated within the floors of major preglacial valleys within the SWSS, unit 1 of the Empress Group is interpreted to be fluvial sediment sourced from rivers flowing from the Cordillera that cut northeast across Alberta before entering the study area (Stalker, 1968). This interpretation is supported by the presence of clasts primarily of chert and quartzite, indicative of this source (Cummings et al., 2012). Unlike larger clasts, sand of EMP-1 probably had multiple sources and is equally likely to have been sourced in the same way as gravels (from the Cordillera) or from the underlying bedrock within these valleys.

Based on previous work within southwestern Saskatchewan (Stalker, 1968) and eastern Alberta (Andriashek and Fenton, 1989) the silt and clay of EMP-2 are hypothesised to be the product of one of two mechanisms: **1.** sediments deposited by rivers in preglacial valleys or; **2.** sediments deposited in lakes formed by the blockage of these valleys by glacial ice. While evidence that directly supports only one of these hypothesises is minimal, silt and clay are described as laminated to finely bedded, thus a lacustrine origin is tentatively favoured. Additionally within the eastern section of the North Battleford Valley, EMP-2 is deposited as a thick deposit covering a vast area (55 km^2), supporting the idea of an extensive glacial lake that flooded the entire valley. Conversely within the Tyner Valley, EMP-2 is more sporadic in appearance. This sporadic appearance may be a result of erosion by the glacier responsible for the deposition of the DU-1, a process that has been attributed to the sporadic appearance of this unit in the Sand River area 73L of Alberta (Andriashek and Fenton, 1989) and the lower Red Deer River (Evans and Campbell, 1995) areas of eastern Alberta. EMP-3 is proposed to be of glaciofluvial origin. This interpretation is made based on the presence of granitic clasts, which indicates they have travelled from the Canadian Shield. A fluvial rather than lacustrine origin is favoured as the sediment is restricted to the valley floors (Evans and Campbell, 1995).

6.2.2 Origin of Diamicton Unit 1

DU-1 is interpreted to be till deposited during the earliest recorded glaciation of the region. However as noted by previous studies in the prairies (Andriashek and Fenton, 1989; Christiansen, 1968b), because this basal till unit is found at or near the base of major buried valleys, some of the unit may include diamicton derived from sediment slumping from the valley walls. Local differences in the grain size of the unit are evident. Glacial erosion and incorporation of the underlying sand and gravel of the Empress Group is one explanation for the variance in grain size (Andriashek and Fenton, 1989).

Within the central Tyner Valley, DU-1 also contains discrete interbedded masses of clay, silt and sand. This suggests that the underlying bedrock was eroded and large amounts were also incorporated into DU-1. Other evidence for bedrock erosion and incorporation are seen in driller log descriptions where DU-1 is reported to contain 'ice rafted shales' and 'claystone'. Chrisitansen (1968) also noted an abundance of clay, sand and shale in the clast fraction of the lowermost till within Saskatchewan, and he also attributed this enrichment to the erosion and incorporation of material by an early glaciation in the area. Additionally, the absence of DU-1 in the northern tributary of the Tyner Valley could be the result of fluvial erosion that occurred during the retreat of the ice margin responsible for depositing DU-1. Glacial meltwater could have eroded this till and deposited thick sand and gravel.

6.2.3 Origin of Stratified Unit A

The composition of this unit suggests that its origin varied both spatially throughout the study area and temporally resulting in distinct layers of clays, sands and gravels. The thick, often clayey deposits of this unit that lie at the base of multiple boreholes are interpreted as glaciolacustrine in origin. In locations where these deposits are interbedded with sand and gravel, they are interpreted as having a glaciofluvial origin. Furthermore a glaciofluvial origin is also inferred for the thick deposits of sand and gravel that make up the rest of this unit. Such a conclusion is based on the location of this material being restricted to buried valleys and channels.

It is not known whether this unit was deposited in association with the underlying DU-1 or the overlying DU-2. Due to the high level of preservation of DU-1, it is likely that at least some of SU-A was deposited immediately after DU-1 and thus acted to protect it from erosion during the following non-glacial period.

6.2.4 Origin of Diamicton Unit 2 and 3

DU-2 and 3 are interpreted as tills deposited during an early glaciation of the study area. Due to the compositional similarity of these units and the lack of a substantial thickness of SU-B, it is inferred that they were deposited during the same glacial period. It is not known whether these till units represent separate glacial advances rather than a glacial readvance in relation to DU-1. However, based on the thickness of SU-A it is proposed a significant ice-free period likely followed the deposition of DU-1, and it is thus reasonable to hypothesise that DU-2 and 3 were produced during a later glacial period. It is unknown if DU-2 and 3 were deposited by two separate ice advances or whether they are the product of compositional differences resulting from variation in material source during one single glacial advance/readvance. Along the Lloydminster Channel, Turtlelake Upland and small areas of the Tyner Valley system, sequences of sand and gravel (SU-B), interpreted as glaciofluvial in origin, separate DU-2 from DU-3, indicating that these tills resulted from two separate advances. Additionally the sharp non-gradational contact between DU-2 and 3 in many boreholes could also represent a break in deposition, potentially pointing to two glacial advances. Furthermore a significant thickening (up to 64m) of DU-3 is present within the northwest of the SWSS. Based on its slight arcuate form, it is tentatively suggested that this thickening could indicate an ice marginal position was present in this location during the retreat of ice that deposited this unit.

Erosion by glacial ice and incorporation of the underlying bedrock is proposed as the source of the abundant silt and clay in DU-2, a process also proposed for the unit of silty, clay rich tills in the Sand River area 73L of Alberta (Andriashek and Fenton, 1989). In contrast the abundance of quartz and sand in DU-3 indicates a considerable amount of quartz sand was eroded during glaciation, however its origin is less easily determined. Its composition is plausibly the result of material from either: **1.** a distant source of quartz sandstone from the Athabasca Formation or; **2.** a more local source from the underlying SU-A. Andriashek and Fenton (1989) suggest that the latter is responsible for the quartz sand rich composition of lower tills (Bonnyville Formation) within the Sand River area 73L of Alberta, though no such conclusion can be drawn in this study and further investigation of this unit's composition is needed.

6. 2. 5 Origin of Stratified Unit C

This unit commonly occurs within areas that would have been low topography when DU-3 was the surficial stratigraphic unit (Map Sheet 3) (though SU-C also occurs on palaeo-slopes in bedrock). It is thus tentatively suggested that these areas contained proglacial lakes and streams that formed as part of a developing drainage network (Andriashek and Fenton, 1989). Therefore silt and clay of this unit are interpreted as glaciolaustrine, deposited in glacial lakes that formed during the advance of glacier responsible for depositing DU-4. It is likely that most of the sediment deposited within these proglacial lakes was probably derived from meltwater that ponded against an ice margin. The arcuate shape of silt and clay deposits likely reflects this.

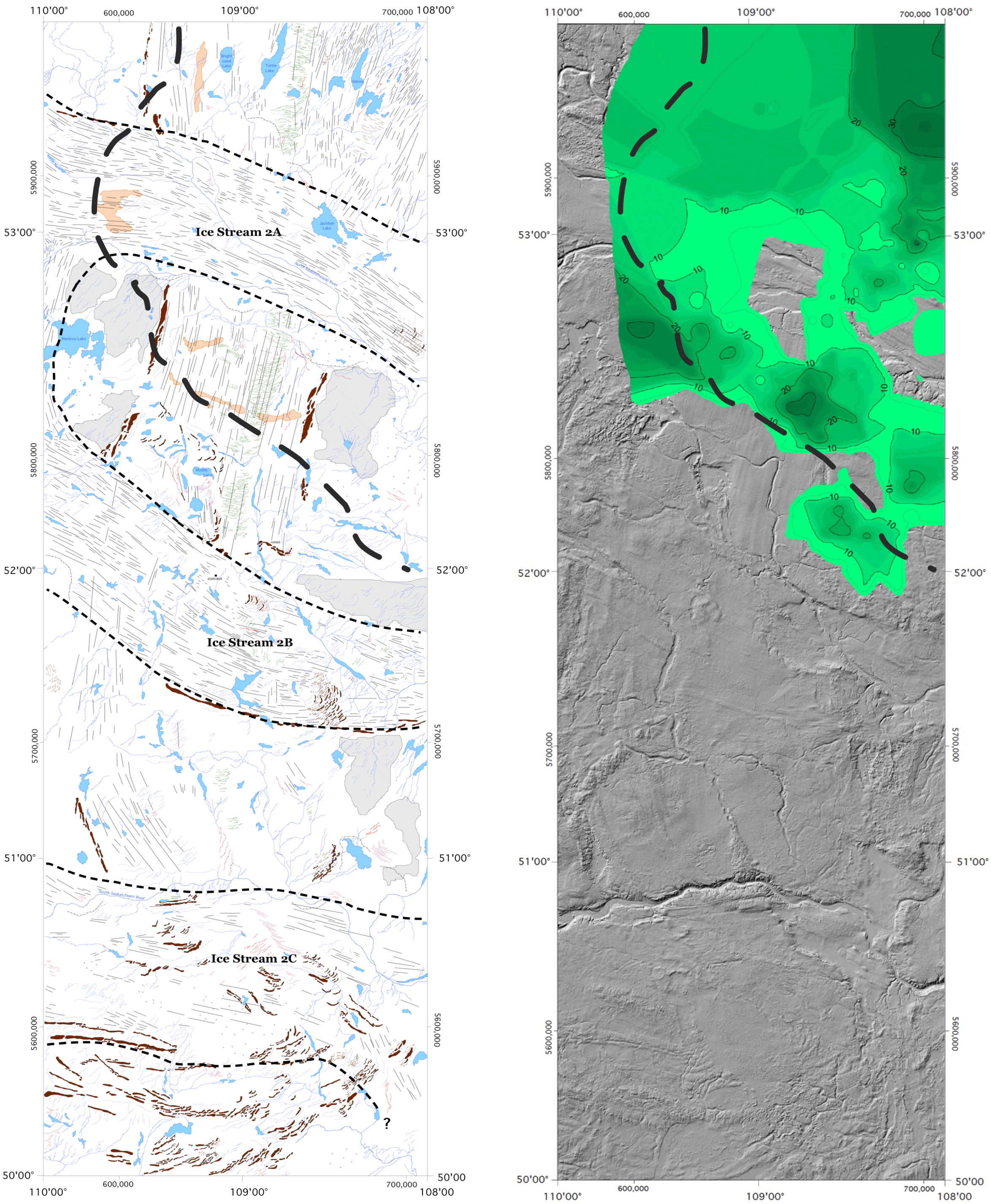


Figure 6.8: *Geomorphic and sedimentary imprint of an ice advance responsible for the deposition of DU-4. Note the clear correlation between the thick lobe deposit of DU-4 and the lobe form of overridden moraine (orange outlines) highlighted by a black dashed line.*

6.2.6 Origin of Diamicton Unit 4 and 5

DU-4 and 5 are interpreted as tills deposited in quick succession during the same glaciation. Evidence to support this is the lack of substantial intertill deposits separating the two units. These tills are interpreted as either glacial advance or re-advance tills deposited within the third recorded glaciation.

The arcuate wedge of DU-4 (and SU-C) in the central and south of the study area indicates an ice margin formed in this region of the SWSS. This proposal is supported by several pieces of evidence. Firstly, the thickening of the DU-4 in this area is consistent with models of glacial sub-marginal till thickening (Boulton, 1996a,b). Secondly, a series of overridden ridges are also present (Fig 6.8) that coincide with recorded terminus thickening. These features are plausibly terminal moraines deposited and overridden by later glaciations. Finally, thick sequences of silt, clay, sand and gravel in SU-C, which are present south of these overridden ridges and till extent, exhibit much similarity to features deposited in a kame terrace formed along ice fronts in contemporary glacial settings (e.g. Bennett et al., 2000; Livingstone et al., 2010) and also to sediments interpreted as kame terrace deposits associated with an overridden ice margin in the Sand River area 73L of Alberta (Andriashek and Fenton, 1989).

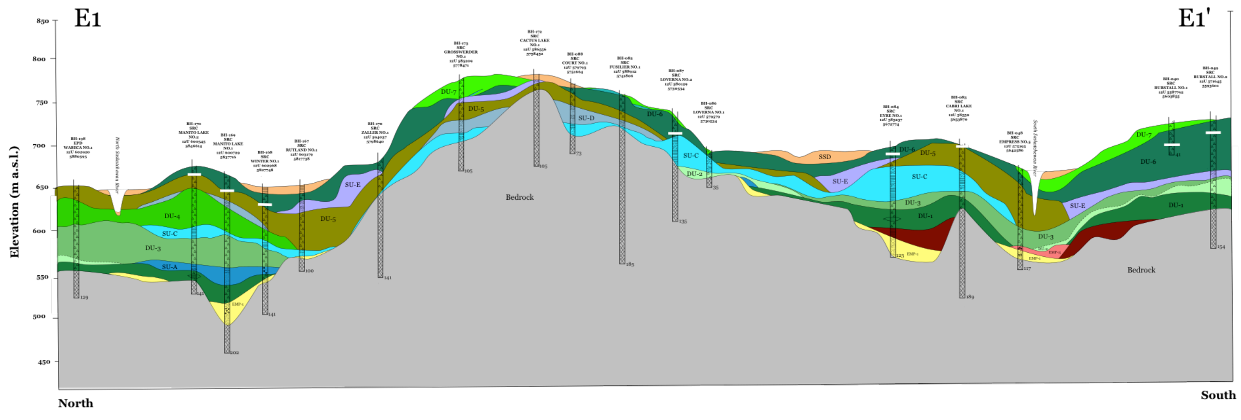
Like DU-4 this unit also shows a thickening in a southwesterly direction, further supporting the idea that these till units were deposited by ice flowing in the same direction during a single glacial period. However DU-5 in contrast to DU-4 is recognised within the southwest of the study area. Consequently both tills likely record flow from the northwest to southwest. In some areas a thin sand and gravel intertill deposit (SU-D) is recorded; due to the limited spatial extent of this deposit, and the fact that it is restricted to areas where DU-4 is recorded, this unit is interpreted as a glaciofluvial deposit formed in association with ice that deposited DU-4. Furthermore in addition to this significant southwesterly thickening in DU-5, the unit also thickens locally within the Battleford buried valley system (and thins immediately south of the system). This localised thickening suggests that pre-existing topography and valley sediment fills may act as a control to the subglacial generation of sediment.

6.2.7 Origin of Stratified Unit E

Not all the sediment in this unit is considered to have been deposited as a result of the same process, nor was it all deposited at the same time and therefore its deposition is, like SU-A, inferred to have been spatially and temporally transgressive. Within the eastern part of the Battleford buried valley system and north of the Western Hills Upland, a silty and clay rich lower layer of SU-A is present. This sediment, unlike the majority of this unit, is oxidised. Based on the restricted distribution of this layer of the unit and its oxidised nature, it is proposed that this sediment was deposited in close succession to DU-5/4 in two large proglacial lakes. A widespread unoxidised layer of sand overlies both the oxidised silt and clay and the oxidised till of the DU-5 and 4. This layer makes up the majority of this unit and is proposed to have been deposited glaciofluvially in association with ice that deposited DU-6.

6.2.8 Boulder Pavement

Within several parts of the study area the basal contact of DU-6 is commonly marked by a boulder layer. Based on its wide distribution and its occurrence at the contact between DU-6 and 7, this concentration of boulders is interpreted as a boulder pavement. Christiansen (1968a) also reports the presence of a pavement of 'faceted and striated' boulders present locally within southern Saskatchewan, which separate a 'sandy, olive brown, stained till' from an overlying 'soft, massive, unstained. surficial till'. Likewise Meneley (1964) reports that a boulder pavement occurs in the Melford area 73A of Saskatchewan. Comparable boulder concentrations have also been reported at similar stratigraphic positions within western Manitoba (Klassen, 1979) and southeastern Alberta (Evans et al., 2012). However it should be noted that while it is likely that some of these boulder pavements were deposited synchronously, a lack of regionally integrated borehole records and poor chronostratigraphic controls means that no correlations can be made between such boulder concentrations.



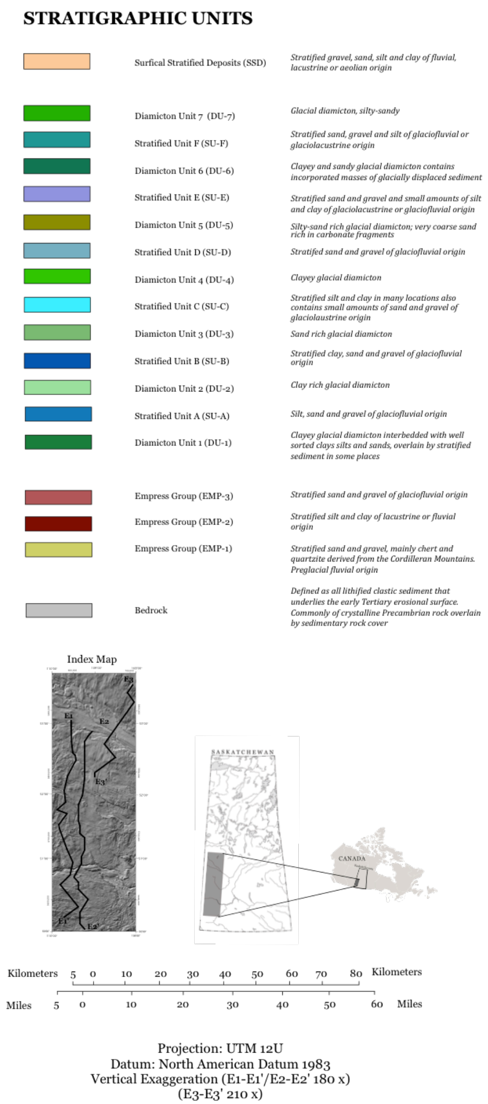


Figure 6.9: *Extract of cross sectional map (see Map Sheet 3 for full image) showing the location of the boulder pavement recorded within DU-6 and/or separating DU-5/stratified sediment underlying DU-6 are represented as white dashes. Note the occurrence of this boulder pavement situated in the middle of DU-6 in the south of the study area and at the contact between DU-6 and the underlying sediment in the north of the study area.*

A variety of mechanisms have been postulated to explain the formation of a boulder pavement (see Clark, 1991), however the clustering of boulders between DU-5 and 6 (overlaying SU-E in some lowland areas) within the centre of the study area (Fig 6.9) and the occurrence of this pavement well within DU-6 at the southern end (Fig 6.9) potentially suggests it has been formed as a lag deposit by excavational deformation at the base of the deforming layer in association with advance and retreat phases of glacial ice (Boulton, 1996a). According to this model, downward movement of the base of a deforming layer into an older till unit will result in the preferential mobilisation of fine material and the accumulation of large particles, which resist entrainment. Thus the occurrence of this boulder concentration in the middle of DU-6 suggests minimal erosion and a large volume of advance phase till overlain (marked by the boulder pavement) by a thick retreat till sequence. Conversely the occurrence of this boulder pavement overlying stratified sediments and/or DU-5 is testament to a zone of strong erosion with a till derived from the retreat phase only.

6.2.9 Origin of Diamicton Unit 6

The diamicton of this unit is proposed to be till deposited by ice flowing north-south through the SWSS. This proposal is made based on the orientation of glacially streamlined features aligned north-south which are composed of this unit. A series of terminal moraines occur at the southernmost occurrence of this till unit. A substantial thickening of this till is seen across the study area (Fig 6.10). The unit increases from 11 m in the north of the SWSS to a maximum thickness of 30 m in the south where it forms large terminal moraines. In contrast the till is very thin within the central lineated corridor especially between the overlying flow paths of Ice Streams 2B and 2C, potentially indicative of rapid streaming ice (Piotrowski and Kraus, 1997; Englehardt and Kamb, 1998). This interpretation is reaffirmed by the convergent corridor of MSGLs that are eroded into the top of and are spatially convergent with the thin zone of this unit. Additionally this unit also seems to vary in thickness in association with the Battleford and Tyner buried valley systems (Fig 6.10). Within these valleys DU-6 thickens and immediately south of the Tyner Valley the unit thins considerably, suggesting that pre-glacial topography and pre-existing valley fills were crucial to subglacial sediment generation. Within the east of Corridor 2A and in the west of Corridor 2C stratified sediment lies locally (SU-F) between

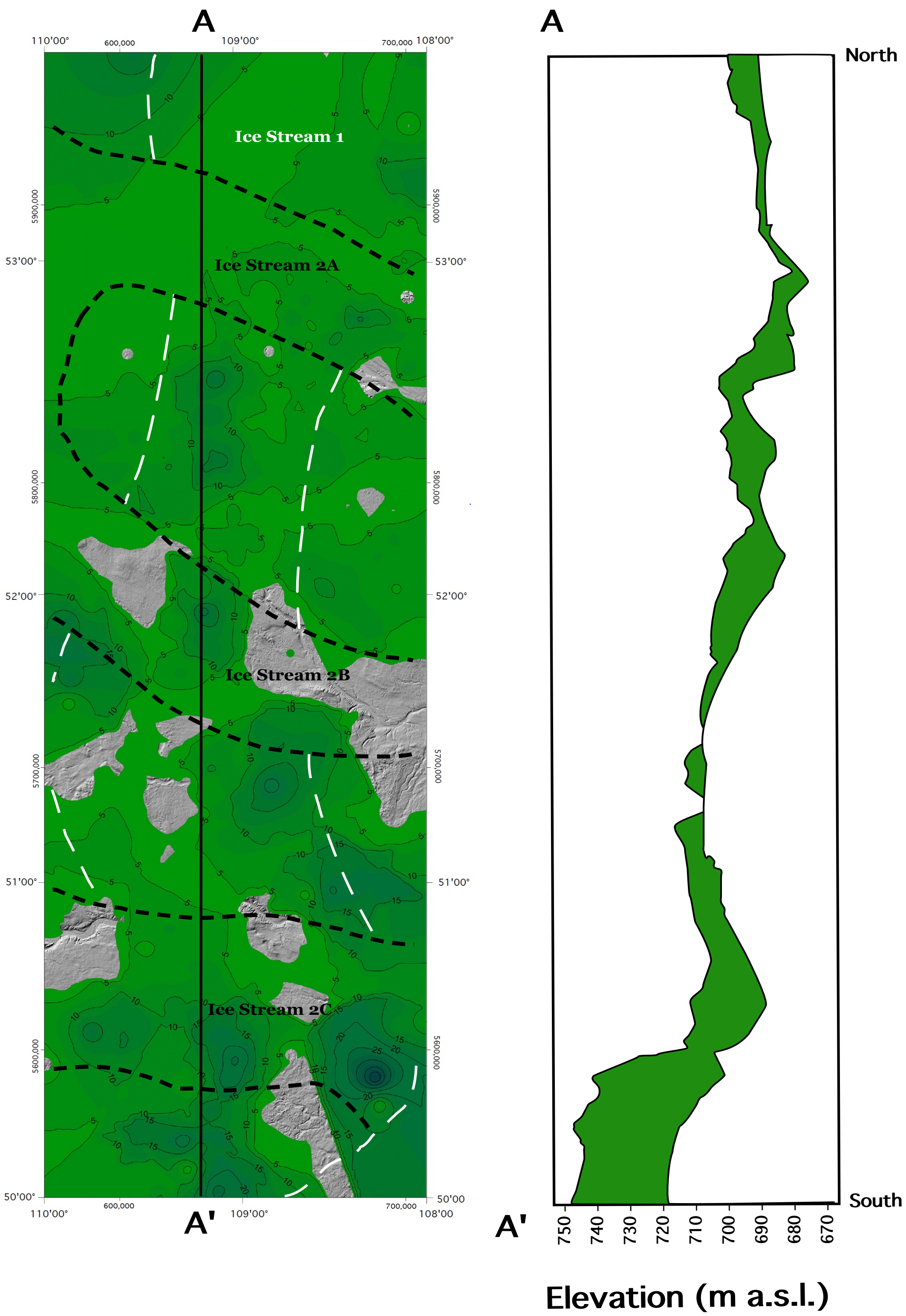


Figure 6.10: *2D isopach map and transect through the SWSS showing the thickness of diamicton comprising DU-6. Dashed lines demarcate flow paths of ice streams. Note the marginal thickening in the south of the study area and the thinness of the unit in the centre of Ice Stream 1's trunk. Due to the irregular distribution of borehole data across the study area and, the extrapolation of contour lines based on such points within rockworks, isopach contour lines should be viewed as an estimate of sediment thickness.*

DU-6 and 7. Glaciofluvial or glaciolacustrine sediment was deposited on top of DU-6 during the later phase of ice responsible for depositing DU-6 or during the early phase of the advance of ice that deposited DU-7.

6.2.10 Origin of Diamicton Unit 7

DU-7 is interpreted as till deposited by three ice streams that advanced from the northeast of Alberta (Fig 6.11). The distribution of glacially streamlined features on the surface of this unit indicates ice was fast flowing and probably confined by the topography of the region. This is particularly evident in the northernmost of the three corridors where DU-7 and NNW-SSE orientated glacial lineations are present. The ice that deposited this till unit was also strongly erosive and displaced large parts of the underlying units via glacial thrusting. This is particularly evident where large ice thrust ridges are present within Corridor 2B (Ó Cofaigh et al., 2010). It is also inferred that the lack of DU-6 in the easternmost section of Ice Stream 2B can be attributed to limited deposition during ice advance and retreat and strong erosion in this location, which has removed any contemporaneous till. Due to a lack of boreholes the extent of this unit in the west North Battleford Valley is unknown, however based on the clear geomorphological evidence (MSGL's and ice stream shear moraine ridges) of ice stream actively, it is highly likely that DU-7 is present in this region, though to confirm this further borehole data would be needed. Within the east of Ice Stream 2A and in the west of Ice Stream 2C, stratified sediment lies locally (SU-F) between DU-6 and 7.

6.2.11 Origin of Surficial Stratified Deposits

SSD are interpreted as a wide array of deglacial lacustrine, outwash, and ice-contact sediments and postglacial alluvium, colluvium, aeolian and landslide deposits. The most significant of these deposits are glaciolacustrine deposits, which in places reach ~100 m, especially in the south of the study area. These were likely deposited in glacial lakes impounded by the ice margin as it retreated downslope towards the northeast (Christiansen, 1979).

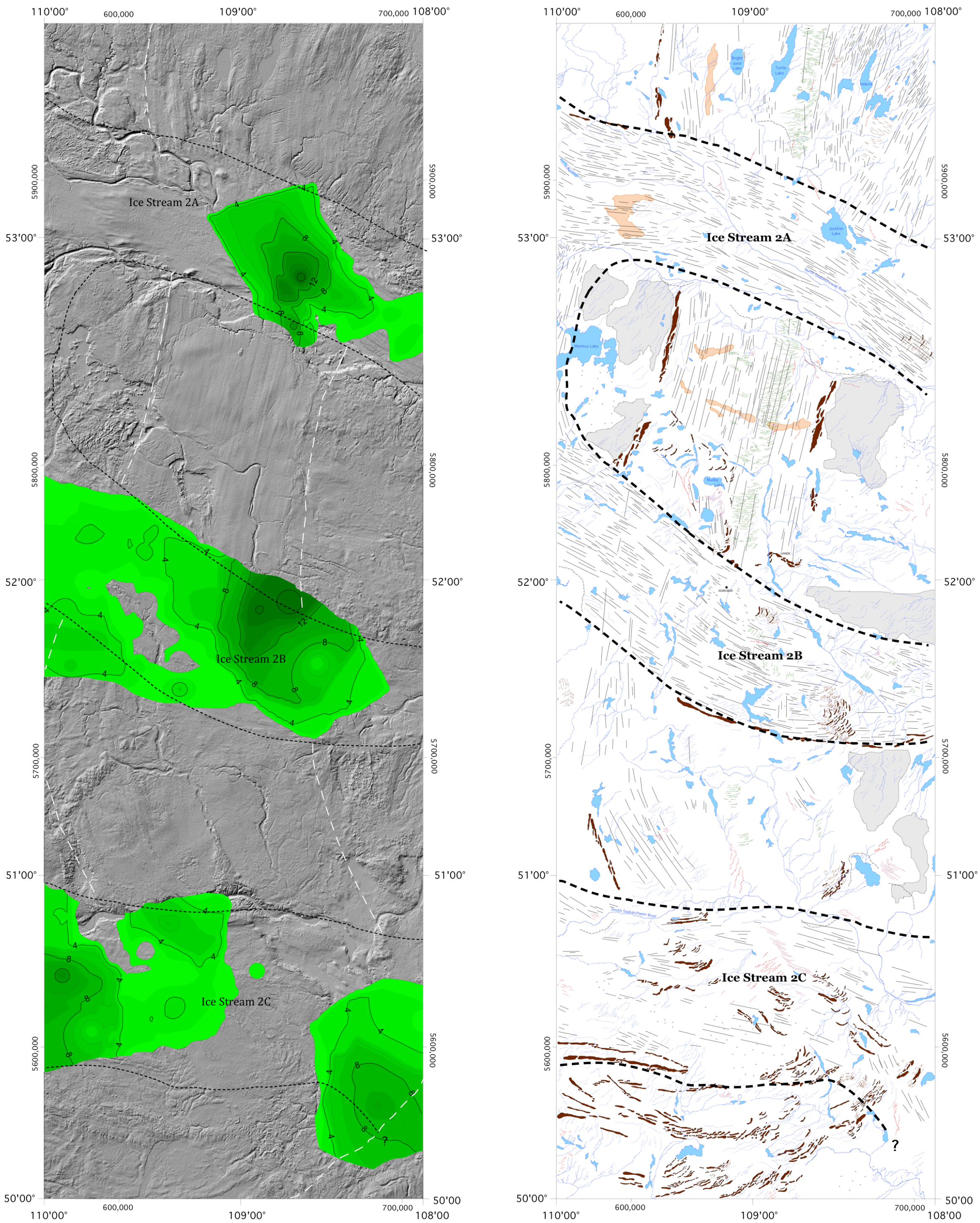


Figure 6.11: *Sedimentary and geomorphic imprint of Ice Streams 2A, B and C. Dashed lines demarcate flow paths of ice streams.*

6.3 Summary

This chapter has presented a sediment-landform model within the SWSS consistent with the evolution of multiple ice stream systems. The following key interpretations should be noted:

1. The smoothed cross cutting Corridors 1 and 2A, B and C are interpreted as zones of former subglacial streamlining within the southwest margin of the LIS, produced by ice streams flowing at higher velocities than surrounding ice.
2. The spatial relationship of these Ice Streams 1 and 2A, B and C indicate that a major change from southward to southeastward trending ice flow occurred in this region.
3. The juxtaposition of CSRs with lineations and ice thrust features provides strong evidence that suggests Ice Streams 2A, B and C underwent surging.
4. Within the preglacial system, preglacial sediment was deposited within valleys by rivers flowing from the northeast.
5. Upon glaciation of the study area, the deposition of 7 till units record multiple stages of ice advance and retreat, which resulted in the erosion, thrusting, incorporation and deposition of thick layers of sediment.
6. Deglaciation is recorded by the presence of thin surficial stratified deposits consistent with the development of glacial lakes impounded by an ice margin retreating towards the northeast.

7. Discussion

This chapter provides a synthesis of Chapters 4, 5 and 6, drawing links between the contents of each chapter and discussing the wider implications of this thesis for palaeo-ice stream dynamics and regional till architecture. The depositional history of the SWSS and the extent of till emplaced during ice stream operation through space and time are discussed first. Within this section an attempt is also made to correlate stratigraphic units with those previously reported within Saskatchewan and in neighbouring Alberta and Manitoba. Building upon this the latter part of this chapter evaluates the significance of till depositional patterns in the context of theoretical models of regional till architecture (Boulton 1996a,b). Finally this chapter will evaluate the methodology used in this study and provide recommendations for future work.

7.1 Depositional history and regional correlations

One of the two main purposes of this study has been to differentiate and characterise lithologic properties of the Quaternary sediments within the SWSS for the purpose of reconstructing the region's depositional history. Interpretations of till and stratified intertill units indicates that ice advanced over the study area 7 times, each time depositing a uniquely different till. However due to an absence of chronological constraints from boreholes it is not known with certainty if units represent separate glacial advances or readvances during the same glacial period. Based on the thickness of stratified intertill units and the presence and/or absence of weathered and oxidised zones, the possibility of some tills representing readvance events or, ice flow directional switches due to shifting ice stream locations within the LIS are suggested. However, dating of these deposits is needed to increase the confidence with which these conclusions are reached. Below, the depositional history of the region is presented in three sections, preglacial, Pre-Late Wisconsinan, and Late Wisconsinan history. Due to the lack of chronological constraints no further division of the depositional history is given, nor is the timing of Pre-Late Wisconsinan glaciations assigned (see Section 7.1.4 for further age estimates based on regional correlations of stratigraphy). A summary of this depositional history is outlined in Figure 7.1.

7.1.1 Preglacial history

Prior to glaciation two major drainage systems, the Battleford and Tyner, developed within the study area. During their active phases it is very likely that both of these rivers had their source in the Cordillera, moving large amounts of material eastwards and depositing quartzitic and chert gravel and sand (EMP-1) on the broad Battleford and Tyner valley floors. However as discussed in section 6.2.1 there is probably more than one source for the EMP-1. Due to the presence of sand composed of quartz and dark chert it is proposed that bedrock was also eroded locally within the study area and deposited in conjunction with gravel sourced from the Cordillera. This depositional model is consistent with previous investigations of basal sands and gravels that occurred within preglacial valleys in southern Saskatchewan and adjoining areas of Alberta (referred to as 'Saskatchewan Gravels and Sands' by Stalker, 1967 and the 'Empress Group, Basal Gravels' by Whitaker and Christiansen, 1972). These deposits thus represent a drainage system that developed roughly through the second half of the Tertiary (there is at present no more definite age assigned to the deposition of EMP-1) and was active until the onset of glaciation ended the prolonged episode of deposition (Whitaker and Christiansen, 1972).

7.1.2 Pre-Late Wisconsinan history

7.1.2.1 Glacial Event 1

The advance of the first LIS ended deposition of EMP-1 (Fig 7.1). Drainage within the Battleford and Tyner valley systems would have been progressively blocked, and diverted southwards. As a direct result of the blockage caused by ice advance, a series of ice-ponded lakes developed. (Fig 7.1). Within these lakes suspended sediment (EMP-2) would have been deposited either overlying coarser basal sand and gravel or in areas where EMP-1 was eroded, sediment would have been deposited on the bedrock. It is also likely that due to a series of ice-ponded lakes developing, water may have produced newly eroded channels; a process alluded to by Christiansen (1968) within southern Saskatchewan. Glaciofluvial sand and gravel of EMP-3 was then deposited either on top of EMP-2 or in some areas on the underlying bedrock (Fig 7.1). It is proposed that as this first ice sheet advanced it eroded the upper surface of exposed bedrock in many places (Fig 7.1). This assertion is made based on the clay rich composition of this till, which likely results from erosion of the underlying mudstone and siltstone. Due to the confinement of this till (DU-1) within preglacial valleys, it is proposed that ice during this period may have taken the form of small tributary lobes or valley glaciers. However, it is also plausible that ice was significantly more extensive and also occupied upland areas but that the evidence was subsequently eroded.

Overlying DU-1 in some regions is a minor layer of clay (SU-A). It is suggested that this material records the development of proglacial lakes against the retreating ice margin (Fig 7.1). The majority of SU-A, however, consists of glaciofluvial sediment that was deposited and then formed a protective layer over DU-1 during the latter phase of this glacial event in the main Battleford and Tyner valley systems. Interestingly within small areas, especially within the northern tributary of the Tyner Valley, DU-1 is absent and/or sporadic. It is proposed that these areas were exposed and later eroded during the following nonglacial period, removing any DU-1 that may have been deposited.

7.1.2.2 Glacial Event 2

Glaciofluvial material (SU-A) was deposited over much of the SWSS either during the latter phases of Glacial Event 1 or in the early stages of Glacial Event 2. A thick layer of till (DU-2) was then deposited (Fig 7.1). This till displays an abundance of quartz and hence it is proposed this could be the result of erosion of sandstone of the Athabasca Formation to the northeast of the SWSS, but could also result from erosion of local sources within the swath itself (the underlying SU-A) (Andriashek and Fenton, 1989). Furthermore the clay rich nature of DU-2 is likely the result of erosion and incorporation of the underlying bedrock (claystone and siltstone) (Fig 7.1).

Overlying DU-2 but occurring below DU-3 is a thin layer of stratified sediment (SU-B) composed mainly of sands and some gravels. The presence of this unit appearing sporadically but in several regions of the SWSS suggests that more than one ice advance occurred during this glacial event. Following the deposition of SU-B a glacial readvance then extended throughout the study area during which DU-3 was deposited. During retreat of the ice responsible for the deposition of DU-3 this ice margin was present in the North Battleford Valley for a significant amount of time. As discussed in section 6.2.4 a large thickening of DU-3 is present in this valley. It is thus tentatively proposed that in this area, thickening represents an overridden glacial margin possibly resulting from a small ice readvance or marginal stagnation during retreat. Furthermore, based on the orientation and arcuate form of this overridden margin, it is suggested that net advection of till was from the northeast, indicating that ice flowed in a northeast to southwest direction during this glacial event.

As part of the later stages of Glacial Event 2, the surface of DU-3 was eroded either by abundant meltwater and/or by poorly defined regional drainage. This is exemplified within the North Battleford Valley where substantial erosion of DU-3 within the valley's eastern section has occurred. The upper parts of DU-3 are also weathered; while this is not widespread it indicates that in some areas the surface of this unit was either only exposed for a very limited amount of time and/or was weathered and later eroded. Sand and gravel (SU-C) were deposited as this drainage system developed on the surface of DU-3 (Fig 7.1).

7.1.2.3 Glacial Event 3

Prior to the Glacial Event 3 proglacial lakes were present in the study area (Fig 7.1). Silt and clay of SU-C were deposited in these lakes in the eastern sections of the North Battleford Valley and Lloydminster Channels. The lakes formed in much the same way as those during Glacial Event 1 as a result of meltwater flowing off the ice and by blockage of the regional northeastern drainage by the advancing glacier. The arcuate distribution of thick silt and clay deposits of SU-C most likely demarcates the marginal zone of this glaciation or at least marks the marginal zone of a temporary ice margin. The former proposal is favoured here for two reasons. Firstly, in addition to SU-C being absent in the southwest of the study area, DU-4 is also absent and displays a similar lobate distribution confined to the northeast of the study area. Secondly, DU-4 and SU-C both thicken considerably towards this lobate margin. Based upon models of sub-marginal till thickening, this observation in association with an overridden ridge complex that is proposed (see Section 6.2.1) to be a terminal moraine, which records the location of the ice margin. Due to the thin and localised appearance of SU-C it is proposed that DU-5 was deposited only a short time after DU-4 during the same glacial event. The widespread distribution of DU-5 across the entire study area indicates the ice responsible for its deposition extended across the entirety of the SWSS (Fig 7.1).

Interestingly, within the bordering Sand River area 73L of Alberta, a similar lobate distribution of clay rich till is recorded (Andriashek and Fenton, 1989). This lobate margin could represent an extension of the ice marginal position of DU-4 within Alberta. However as no chronological constraints exist for either of these tills, such a proposal can only be tentatively suggested and further work would be needed to infer if these two tills represent a regional ice marginal position or are the result of separate events and processes. During deglaciation of the ice that deposited DU-5, meltwater eroded a number of channels on the surface, depositing some of the stratified sediment of SU-E either from proglacial lakes or streams. Erosion during the following nonglacial period was minor and a widespread weathering zone developed on the surface of DU-5.

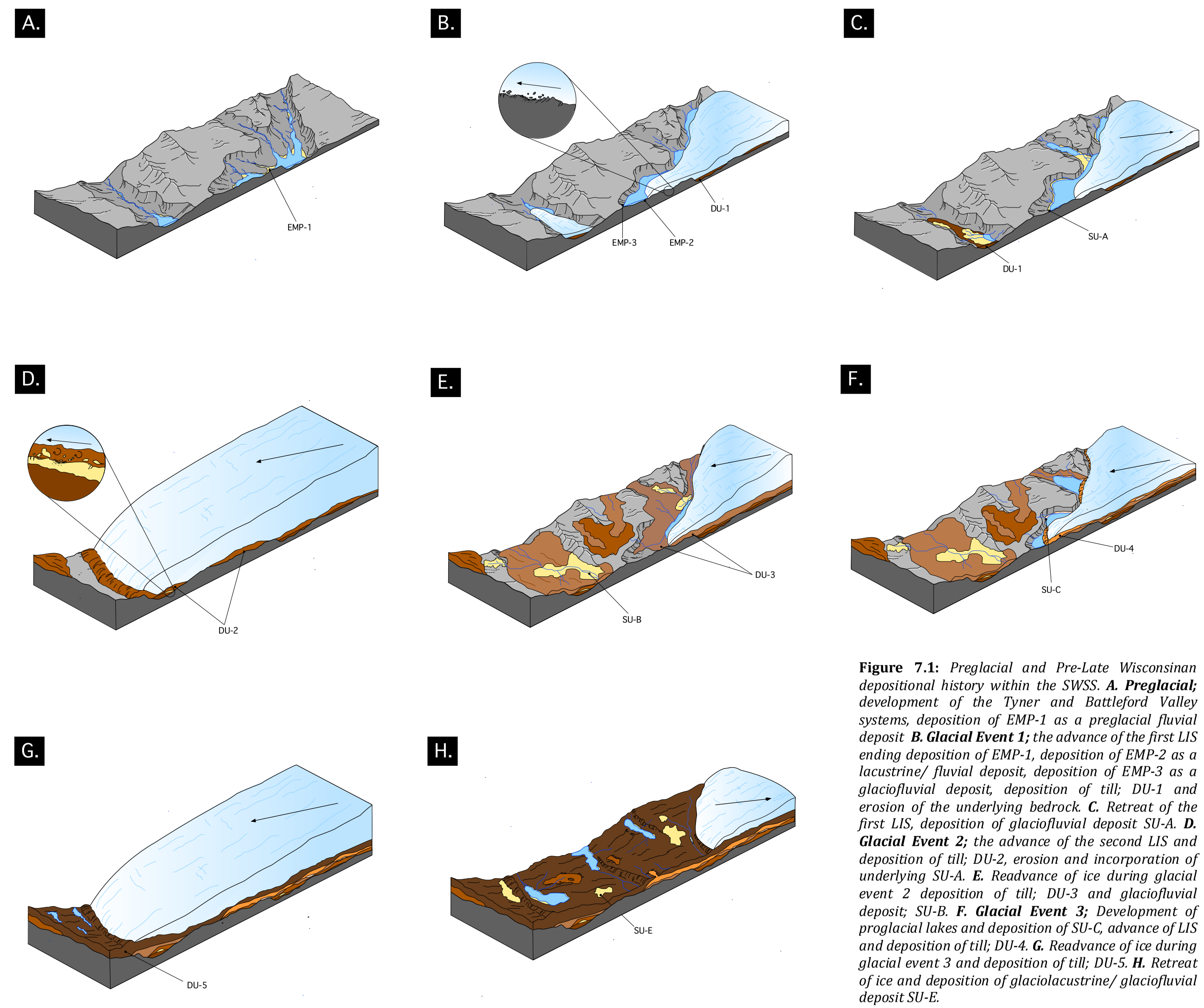


Figure 7.1: *Preglacial and Pre-Late Wisconsinan depositional history within the SWSS.* ***A. Preglacial;*** *development of the Tyner and Battleford Valley systems, deposition of EMP-1 as a preglacial fluvial deposit* ***B. Glacial Event 1;*** *the advance of the first LIS ending deposition of EMP-1, deposition of EMP-2 as a lacustrine/ fluvial deposit, deposition of EMP-3 as a glaciofluvial deposit, deposition of till; DU-1 and erosion of the underlying bedrock.* ***C.*** *Retreat of the first LIS, deposition of glaciofluvial deposit SU-A.* ***D. Glacial Event 2;*** *the advance of the second LIS and deposition of till; DU-2, erosion and incorporation of underlying SU-A.* ***E.*** *Readvance of ice during glacial event 2 deposition of till; DU-3 and glaciofluvial deposit; SU-B.* ***F. Glacial Event 3;*** *Development of proglacial lakes and deposition of SU-C, advance of LIS and deposition of till; DU-4.* ***G.*** *Readvance of ice during glacial event 3 and deposition of till; DU-5.* ***H.*** *Retreat of ice and deposition of glaciolacustrine/ glaciofluvial deposit SU-E.*

7.1.3 Late Wisconsinan history

The final major glacial event to affect the study area left its imprint not only in the stratigraphy but also geomorphologically. This additional data allows a more detailed reconstruction of the history beyond the simplistic advance and retreat phases of earlier events. As discussed above, chronological constraints on the timing of glacial occupation of the study area are limited. However in order to place both geomorphological and sedimentological observations into a chronological context this study will employ the chronology described by Ross et al. (2009) and Ó Cofaigh et al. (2010) based on the reconstruction of LGM deglaciation by Dyke et al. (2003). In order to allow a discussion of the dynamics of ice during the Late Wisconsinan the geomorphological and sedimentary results of this study are placed in a regional context (Fig 7.2). Thus in combination with this chronology the following is an updated and refined reconstruction of the Late Wisconsinan history that builds upon the previous work completed within southeastern Saskatchewan (Ross et al., 2009; Ó Cofaigh et al., 2010) but also integrates reconstructions of ice stream flow within Western Manitoba (Patterson, 1997; Jennings, 2006) and Western Alberta (Andriashek and Fenton, 1989; Evans et al., 2008).

7.1.3.1 Relationship between surficial landscape and sub-surface stratigraphy

At the LGM ice covered Saskatchewan and Alberta and extended into Montana and North Dakota (Dyke et al., 2002; 2003). The well preserved geomorphic ice stream flow sets, discussed in section 6.1.6, represent the dynamic behaviour of this part of the ice sheet during its retreat from this maximum position. Based on the relationship of cross cutting Ice Streams 2A, B and C in relation to the flow path of Ice Stream 1, it is clear they record progressive changes in ice sheet behaviour through the Late Wisconsinan deglaciation from an unconfined 'pure' stream to a series of topographically confined flow lobes (Fig 7.3). It would seem probable that this dynamic flow shift is also recorded within the stratigraphic record. Based on sediment-landform relationships it is proposed that DU-6 was deposited by Ice Stream 1 and DU-7 relates to the activity of Ice Streams 2A, B and C. This interpretation is based on several lines of evidence: **1.** the corresponding narrow distribution of DU-7 within three strips all within smoothed corridors 2A, B and C; **2.** the stratigraphic position of DU-7 which overlays DU-6 in areas where the flow paths of Ice Streams 2A, B and C and Ice Stream 1 are present; **3.** thickening of DU-6 in relation to the terminus of Ice Stream 1.

7.1.3.2 Phase 1: LGM ~18 ^{14}C ka BP- Ice Stream 1 active flow phase

The flow path of Ice Stream 1 can be recognised from the Canadian Shield to the southeastern corner of the SWSS. The corridor of MSGLs that document this palaeo-ice stream can be followed down to the Great Sand Hills Lowland. This indicates that during the LGM (~18 ^{14}C ka BP; Dyke et al., 2002), this ice stream terminated on the northern slope of the Cypress Hills. However based upon the occurrence of MSGLs, that extend outside of the ice stream's terminus (as demarcated by a series of moraines) towards the southeastern corner of the SWSS, at the southwestern corner of the Cypress Hills, it is plausible that Ice Stream 1 may have fed ice to lobes documented further south between the Cypress Hills and Wood Mountain Uplands (Klassen, 1991, 1992, 1994; Ross et al., 2009). At the same time similar fast ice flows were also active within the bordering provinces of Manitoba and Alberta. The James Lobe to the east of the study area was fed by a system in Manitoba and eastern Saskatchewan (Patterson, 1997; Jennings, 2006). Within the CAIS (Central Alberta Ice Stream) and HPIS (High Plains Ice Stream) were active (Fig 7.2). The HPIS, CAIS, James Lobe and Ice Stream 1 all displayed a general N-S flow against the regional slope and without significant deflection crossed preglacial valleys. This flow pattern indicates that ice was thick over the Interior Plains while these ice streams were active and likely resulted from coalescence with Cordilleran ice in the west (Dyke and Prest, 1987; Klassen, 1989; Rains et al., 1999), which would have forced the net movement of flow southwards (Fig 7.2).

The series of transverse moraine ridges (Map Sheet 1) that form an arc extending from the Great Sands Hills Lowland to the base of the Cypress Hills, developed along the regional slope. This depositional context suggests ice thrusting towards the south-southwest (Fig 7.3). This thrusting process was first proposed by Kupsch (1962) in the context of an ice sheet marginal thrust model. However due to their location in the terminal zone of Ice Steam 1, it is clear these features were produced by palaeo-ice stream activity. Ice thrusting by Ice Stream 1 is thus invoked to explain the thick till sequences found in this area. This thrusting is likely a result of glaciotectonic thrusting along the regional slope over a short period of time as opposed to persistent glacial thrusting along the margins by the ice sheet itself (Fig 7.3). This interpretation would also account for the occurrence of several oxidised horizons within DU-6 in the marginal zone of Ice Stream 1.

7.1.3.3 Phase 2: regional flow reorganisation

Based on the chronology proposed by Dyke et al. (2003), by 14 ^{14}C ka BP the LIS had reached its maximum extent and had begun to thin within western Canada. During this period the ice sheet's marginal regime transitioned from a system characterised by steady state flow to one characterised by dynamic temperate ice-marginal conditions (Ross et al., 2009; Evans et al., 2014). During glacial retreat, the ice margin retreated down-drainage, ponding LIS meltwater in proglacial lakes and trapping drainage from the recently deglaciated areas (Mathews, 1974) (Fig 7.3). Stratigraphic evidence observed in this study and sedimentological evidence reported by Ó Cofaigh et al., (2010) indicates that at some stage in its history, Ice Stream 1 terminated in a proglacial lake(s) or ponded area. This ponding and increasing meltwater lubrication created an unstable ice marginal zone.

A major change in flow direction then occurred with ice streams from farther west dissecting Ice Stream 1 (Fig 7.3). This change in flow direction was first alluded to by Clayton and Moran (1982). In their reconstruction they suggest a regional shift from southward trending ice flow to southeastward trending flow after 14 ^{14}C ka BP. In this study this shift is linked to the shutdown of Ice Stream 1 and replacement by Ice Streams 2A, B and C. As a result of ice sheet thinning and increasing meltwater lubrication, Patterson (1997) proposed that the ice stream system feeding the James Lobe (Fig 7.2), situated to the east of the SWSS, started to expand northwestward subsequent to 14 ^{14}C ka BP in its southern lobes. This eastern system expanded and developed into 3 ice stream corridors within the SWSS, which captured subglacial water associated with Ice Stream 1, resulting in the eventual shut down and stagnation of this Ice Stream (Fig 7.3). This change increased fast flow and subglacial erosion within Ice Stream corridors 2A, B and C, while the interstream areas of the system saw little erosion and sedimentation was absent as demonstrated by the lack of DU-7.

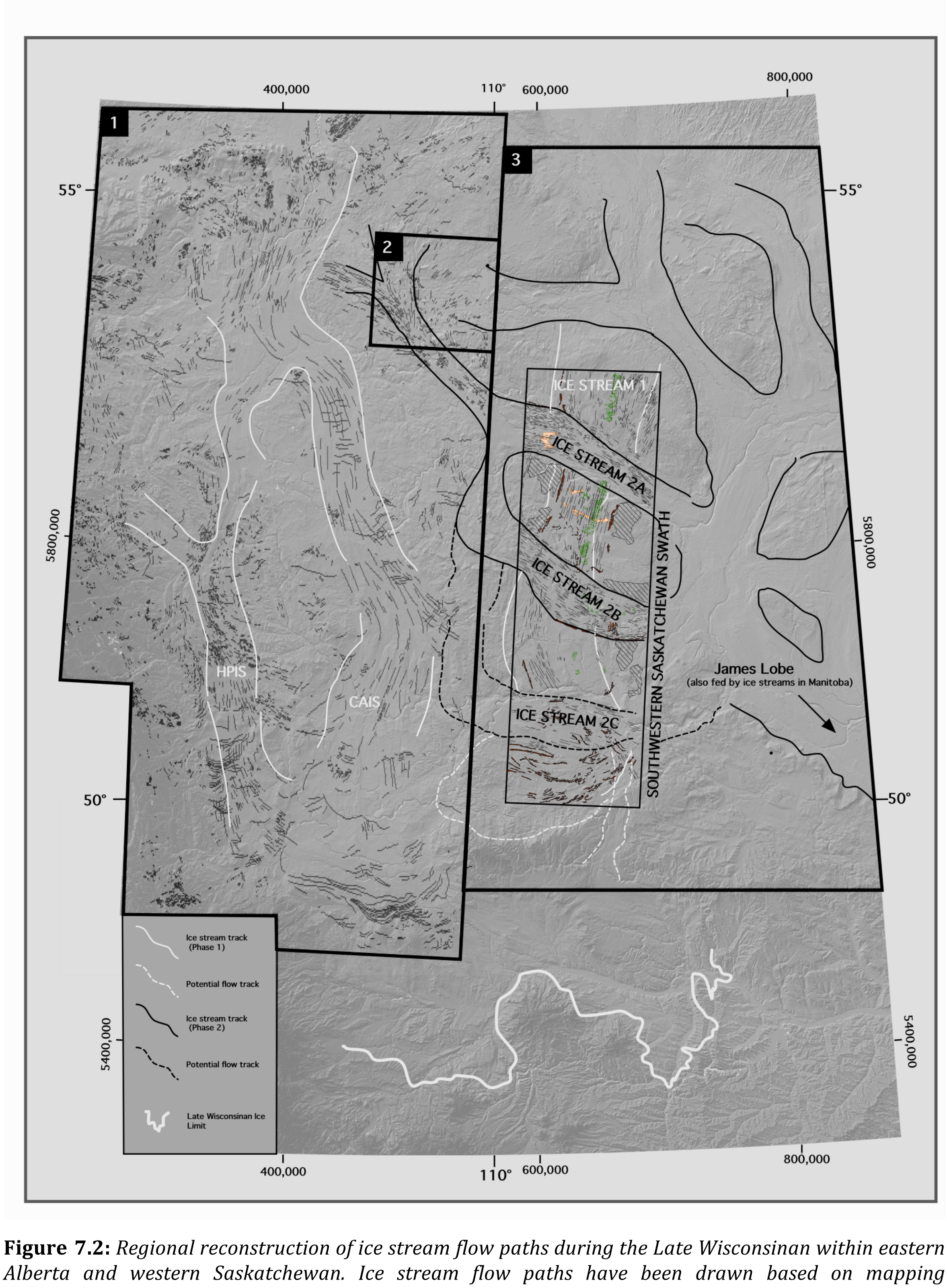


Figure 7.2: *Regional reconstruction of ice stream flow paths during the Late Wisconsinan within eastern Alberta and western Saskatchewan. Ice stream flow paths have been drawn based on mapping undertaken in previous studies **1.** Atkinson et al. (2014) and Evans et al. (2014) **2.** Andriashek and Fenton (1989) **3.** Ross et al. (2009). Late Wisconsinan Maximum extent drawn based on the reconstructed extent proposed by Dyke et al., (2012). Ice streams are separated into flow phases 1 (white ice stream flow paths) and phase 2 (black ice stream flow path). It should be noted that all ice streams in each phase may not have been active simultaneously and there may have been some temporal separation in the exact timing of advance and retreat.*

The stagnation of Ice Stream 1 followed by the progressive migration of the James Lobe onset zone (Patterson, 1997) and associated tributaries explains well the sharp contact between Ice Stream corridors 2A, B and C and Ice Stream 1. The capturing and thus concentration of subglacial water might also explain why Corridors 2A and B display a surge signature. Comparisons with reconstructions of deglaciation of this sector of the LIS (Dyke et al., 2003) implies that the change from unconfined streaming to topographically confined streaming occurred in a couple of thousand years between ~14,500 and 12,500 yrs ^{14}C ka BP. This occurred before the major eastward shift in ice dispersal and prior to complete separation of the Cordilleran and Laurentide Ice Sheets (see Dyke et al., 2003).

7.1.3.4 Phase 3: Ice Stream 2A, B and C active flow phase and Ice Stream 1 stagnation

Within the trunk zone of Ice Stream 1, the good preservation of MSGLs suggests that the main trunk of the ice stream did not undergo active retreat but rather stagnated in situ (Fig 7.3). During this final stage of shut down, linear zones of ice fracturing developed in response to extension and fracturing of the last active flow lobes, with subglacial till injection into these crevasses forming the CSRs (Fig 7.3). This fracturing would have allowed the formation of CSRs within the trunk zone of this ice stream. However, within some small areas at the margins of Ice Stream 1 the subglacial landscape is locally overprinted by younger ice recessional or ice readvance features. Later stage assemblages of features clearly cross cut the subglacial bedforms and comprise a fluted hill-hole pair and esker complex and small lobate moraine ridges. It is proposed that these features demarcate the incursion of a small glacial lobe flowing from the NNW-SSE (Fig 7.3).

In contrast to Ice Stream 1, Ice Streams 2A, B and C are constrained within the low topography of the Battleford and Tyner buried valley systems. The location of the onset zone of these ice streams is unknown. Comparison of geomorphological mapping in this study with mapping by Andriashek and Fenton (1989) in the Sand River area 73L of Alberta indicates that at least Corridors 2A and B within the SWSS are a clear continuation of the Lac La Biche Ice Stream that flowed NNE-SSW through the Sand River area and split into multiple lobes within Saskatchewan. A source within central Alberta is further supported by the low carbonate concentrations of surficial tills that are found along the Ice Streams 2A, B and C relative to Ice Stream 1, which are consistent with a provenance from carbonate poor Alberta (Ross et al., 2009). Their location is also consistent with the proposed palaeo-ice stream path to the southeast that ended with the James Lobe (Fig 7.2). All three lobes' flow paths can be traced as tributaries consistent with palaeo-ice stream criteria of tributary-type flow patterns and convergent flow patterns. The fields of attenuated bedforms and their large (>10:1) length to width ratios suggest rapid flow speeds occurred along the Ice Stream corridors 2A, B and C towards the southeast. This is especially clear within Ice Stream 2A where MSGLs are highly elongated, indicative of high flow velocity (cf. Dyke and Morris, 1988; Clark 1993; Wellner et al., 2001; Ó Cofaigh et al., 2002; Briner, 2007).

These corridors were bound by the surrounding higher topography, but the system within these topographically defined areas was dynamic and shifts in lateral margins and overprinting of some assemblages are evidence of this instability. It is evident that the landform-sediment assemblages recorded in these ice stream corridors also exhibits all the characteristics (with the exception of concertina eskers) of the surging glacier landsystem model. While the cause of such surging is unknown it is likely that the vast area of deformable substrate within Western Canadian Sedimentary Basin (Mathews, 1974; Boulton and Jones, 1979) would have facilitated surging activity. The low yield stresses and high pore water pressure associated with these sediment beds would have resulted in a low surface gradient ice cover, which as a result of its characteristics would have been prone to abrupt flow change (Mattews, 1974; Fisher et al, 1985). This geological setting would have provided ideal conditions for ice stream surging. Other conditions that may have promoted ice marginal instability in the form of surging, include a thin low-gradient ice margin that would have allowed for cold-based conditions. Surging behaviour was not isolated to the Saskatchewan swath, but also characterised ice stream behaviour closer to the onset zone of this ice streams and within its marginal zone. Within the Lac La Biche area of this ice stream system Evans and Rea (1999) report a surge signature, and in central Manitoba Adams (2009) describes a surge signature associated with the Weyburn Lobe.

Such instability in these ice stream corridors has important implications for the Quaternary history of the Western Plains. The position of this assemblage within these streaming corridors, strongly suggests that surging affected this margin of the LIS during its recession from southwestern Saskatchewan. Furthermore in combination with the evidence from Ice Stream 1 and work within other areas, the Prairies provide evidence for ice-marginal surging, active ice-margin retreat and in situ stagnation. For example, diachronous activities of the LIS have been documented by Stalker (1973) who suggested a glacier surge to produce the Innisfail Lobe, during the Late Wisconsinan, within the Frank Lake area of Alberta, which contrasts with the active sequential retreat of the same lobe further south. Additionally, the landsystem evidence of this surge and that associated with the Lac La Biche area, SWSS and Weyburn Lobe appears to contrast with evidence of fast flow of both the adjacent CAIS (Evans et al., 2014), where well-defined recessional push moraines are largely absent and also the HPIS, where recessional moraines record progressive ice stream retreat (Evans et al., 2008). These observations suggest the possibility of periodic instability within this sector of the ice sheet during the Late Wisconsinan.

7.1.3.5 Late Wisconsinan regional synthesis

Figure 7.2 is a regional reconstruction of interpreted palaeo-ice stream flow paths for the Western Canadian Prairies that integrates the observations of this study with existing reconstructions from Alberta and Manitoba. It indicates that this sector of the LIS was characterised by a complex series of flow events during deglaciation from the LGM limit. Such marked spatial and temporal shifts in flow trajectory suggest that streaming may have been transitory in many cases and did not reach steady state. This is consistent with both reconstructions from the other sectors of the LIS (e.g. Boulton and Clark, 1990; De Angelis and Kleman, 2005; Jennings, 2006) and with observations of modern ice streams (Alley and Bindschadler, 2001) and thus, demonstrates that ice streams are an integral component of a complex glacio-dynamic system capable of operating on a variety of scales and interdependencies.

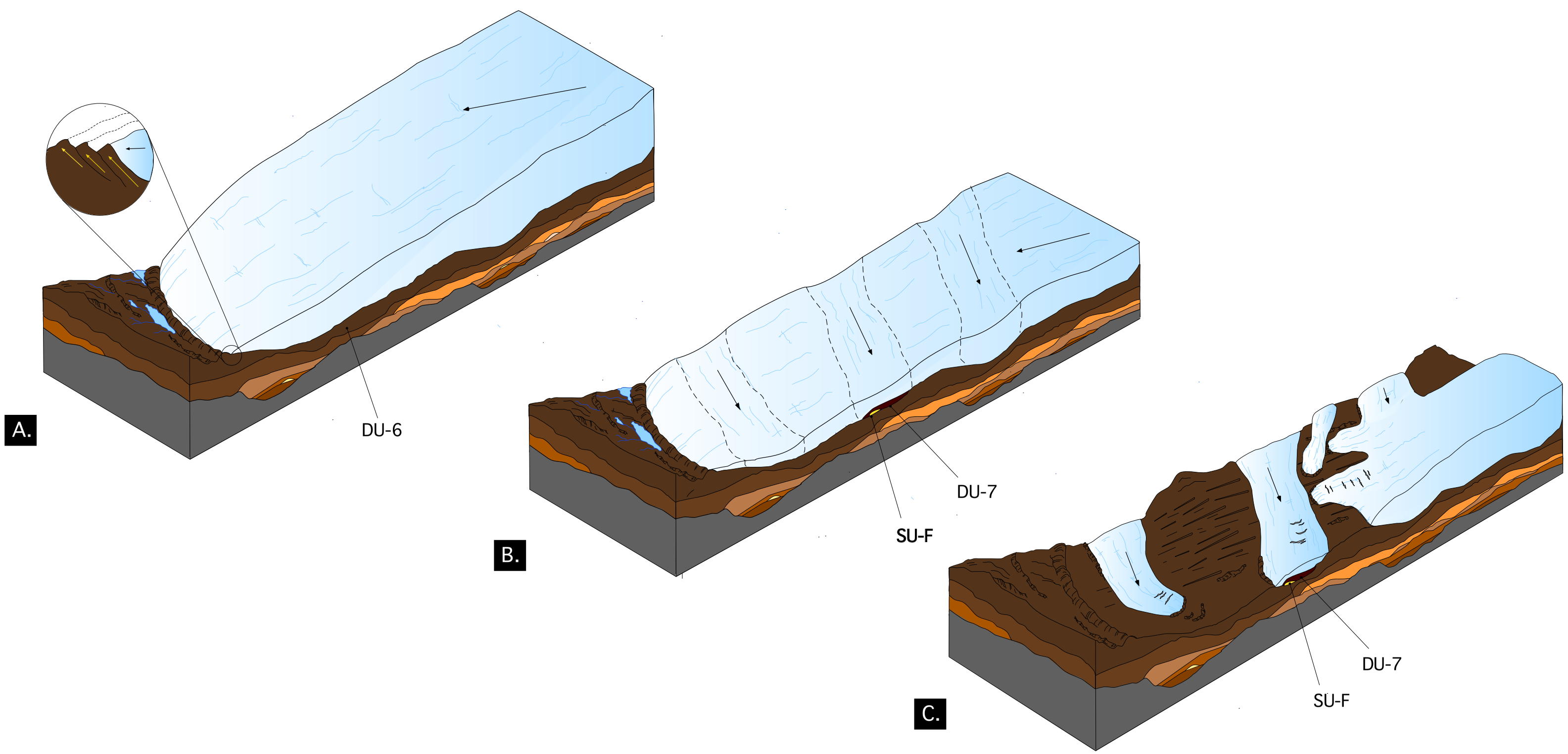


Figure 7.3: Late-Wisconsinan depositional history within the SWSS. **A.** Phase 1: Ice Stream 1 Active Flow Phase; advance of Late Wisconsinan ice sheet and active flow of Ice Stream 1 resulting in the deposition of DU-6. Thick till sequences and a series of moraine ridges developed at the ice stream terminus indicative of glaciotectonic thrusting against the regional slope. **B.** Phase 2: Regional flow reorganisation; after ice reached its maximum extent during the Late Wisconsinan a major change in ice flow direction then occurred with the flow of Ice Streams 2A, B and C from farther west dissecting Ice Stream 1. This new ice stream system expanded and captured subglacial water associated with Ice Stream 1, resulting in the eventual shutdown and stagnation of Ice Stream 1. Flowing from NNE-SSW through the Battleford and Tyner buried valley systems the active flow phases of Ice Stream 2A, B, C was topographically bound. These much thinner ice streams also exhibit a surge signature and based on the occurrence of multiple CSRs it is likely there terminus zones would have bee heavily crevassed. **C.** Phase 3: Ice Stream 2A, B, C active flow phase and Ice Stream 1 stagnation; in situ stagnation preserved moraine ridges, CSR and lineations at the glacier bed. Prior to complete shutdown within the trunk zone of Ice Stream 1 linear zone of ice fracturing developed in response to extension and fracturing of the last active flow lobes, this fracturing allowed the development of CSRs. Incursion of minute glacier lobes flowing from the NNW-SSE into the trunk zone of Ice Stream 1 also occurred leaving small arcuate morainic arcs.

7.1.4 Till: stratigraphy and regional correlation

Repetition and omission of tills due to glacial thrusting, the offsetting and preservation of beds by gravity faulting and major changes of till thickness all occur in the till sequences of the Canadian Prairies. These both confuse identification of individual tills and complicate correlations. As a result of these difficulties only tentative correlations of the lithostratigraphic units in this study with those defined by previous workers in the southern Saskatchewan, southern Manitoba and bordering Alberta can be made. These correlations are displayed in Table 7.1. Estimated ages have also been assigned based on those proposed by Christiansen (1971) though it should be noted that absolute dating has only been undertaken within stratified sediment below the Battleford Formation within southern Saskatchewan. Age assessments can only be stated with confidence for the youngest units (DU-7 and 6) which were deposited during the last glacial and are likely of Late Wisconsinan age.

The criteria used to make these correlations are characterised by differences in the degree of certainty depending on the unit involved. A cross section across Alberta and Saskatchewan would be required to formally correlate these deposits. This would be partially useful to correlate the lower DU-4, 3, 2 and 1. Correlations of these lowest tills are particularly uncertain between the Brocket, Maunsell and Labuma tills and the Sheel, Tee and Largs Formations. Thus to avoid over interpretation, correlation of these units are not proposed beyond generalised groupings (Table 7.1). The correlation with the upper most DU-7 and 6 are made based on their stratigraphic position as the uppermost tills, and their soft, unconsolidated, massive to fissile, less oxidised, nature (compared to underlying tills) and basal boulder concentration. Correlations to DU-5 and 4 are made based on the fractured nature, highly oxidised surface, and higher resistivity than the overlying tills. Correlations to DU-3 and 2 included the use of the oxidised surface, high clay context and low resistivity of the till units. The criteria used to correlate to DU-1 are the high clay content, very low resistivity, high local bedrock content, and the fact that it is the lowest till in the sequence and is confined to buried valleys. Empress Groups units 1-3 are correlated based on stratigraphic position and composition of clasts.

Table 7.1: *Proposed regional correlations (incorporating stratigraphic units identified in this study) and estimated ages (Christiansen, 1972) of lithostratigraphic units compiled. This includes the stratigraphic framework constructed for southern Saskatchewan after Christiansen (1992) and Barendregt et al., (1998). Compiled from: Fulton et al. (1986), Andriashek and Fenton (1989), Klassen (1989) Christiansen (1992) and Barendregt et al., (1998).*

TIME UNITS			ENVIRON-MENT	STRATIGRAPHIC UNITS						
				Saskatoon Area, Saskatchewan (Christiansen, 1992; and Barendregt et al., 1998)			Sand River Area 73L, Alberta (Andriashek and Fenton, 1989)	Southern Alberta (Fulton et al., 1986; Klassen 1989)	Southern Manitoba (Klassen 1979, 1989)	**This study (South-western Saskatch-ewan)**
Quaternary		Holocene	Post-glacial		Stratified Drift		Deglacial and Postglacial Deposits	Surficial Deposits	Surface Deposits	**SSD**
			Deglacial							
	Late Pleistocene	Late Wisconsinan	Glacial	Saskatoon Group	Battleford Fm.	Battleford Till (Upper)	Grand Central Fm.	Buffalo Lake Till	Arran Fm, Zelena Fm. Lennard Fm.	**DU-7**
						Battleford Till (Lower)				**SU-F**
										DU-6
			Proglacial			Weathered Zone	Sand River Fm.	Evil-smelling Band	Unnamed Deposits	**SU-E**
		Middle Wisconsinan	Nonglacial		Floral Fm.					
			Proglacial							
		Early Wisconsinan	Glacial			Floral Upper Till	Marie Creek Fm. (Unit 2)	Cameron Ranch Fm.	Minnedosa Fm	**DU-5**
			Proglacial			Riddel Member		Mitchell Bluff Fm.	Roaring River Clay	**SU-D**
	Early and Middle Pleistocene	Sangamon	Nonglacial							
		Illinoian	Proglacial							
			Glacial			Floral Lower Till	Marie Creek Fm. (Unit 1)	? ↑ Brocket Till Maunsell Till Labuma Till ↓ ?	? ↑ Shell Fm. Tee Lakes Fm. Largs Fm. ↓ ?	**DU-4**
			Proglacial			Weathered Zone	Ethel Lake Fm.			**SU-C**
		Pre-Illinoian	Nonglacial	Sutherland Group	Warman Fm.					
			Proglacial							
			Glacial			Warman Till	Bonnyville Fm. (Unit 2)			**DU-3**
			Proglacial		Dunburn Fm.	Weathered Zone				**SU-B**
			Nonglacial							
			Proglacial							
			Glacial			Dunburn Till	Bonnyville Fm. (Unit 1)			**DU-2**
			Proglacial		Mennon Fm.	Weathered Zone	Muriel Lake Fm.			**SU-A**
			Nonglacial							
			Proglacial							
			Glacial			Mennon Till	Bronson Lake Fm.			**DU-1**
			Proglacial	Empress Group			Empress Unit 3			**EMP-3**
Tertiary	Pliocene		Preglacial				Empress Unit 2			**EMP-2**
							Empress Unit 1			**EMP-1**

7.2 Dynamics of terrestrial terminating ice streams in southwestern Saskatchewan: glacial erosion, transport and deposition

While site specific glacial sedimentology has been placed in a regional context to look at the characteristics of till architecture in relation to ice stream activity and terrestrial moraine and till forming processes, research to date within the Canadian Prairies has lacked comprehensive sub-surface analysis. This study is the first attempt to produce a regional 3D analysis of sub-surface stratigraphy, which allows a large scale investigation of ice sheet till architecture. In the following section the significance of this study will be assessed in the context of till erosional, transportational and depositional processes associated with fast glacier flow. The regional characteristics of the till architecture will then be reconciled with Boulton's (1996a, b) theoretical proposal of till deposition patterns.

Tills that are deposited around the southern margins of the LIS have been widely interpreted as the product of subglacial deformation (e.g. Alley, 1991; Evans and Campbell, 1995) and because subglacial deformation has been closely linked with fast glacier flow (e.g. Boulton and Hindmarsh, 1987; Alley et al., 1986; Dowdeswell et al., 2004), it seems likely that the SWSS tills played a significant role in fast ice flow identified in this study. Conversely widespread sliding has also been reported to have been a significant driver of palaeo-ice stream flow (e.g Piotrowski et al., 2001, 2004; Stokes and Clark, 2003) and the occurrence of large subglacial meltwater channels in the fast ice flow corridors of the SWSS attests to the occurrence of subglacial meltwater during the fast-flow events.

7.2.1 Marginal thickening of subglacial sediments

Evidence of marginal thickening of subglacial sediments is manifest by DU-6 in association with the margins of the southern limits of Ice Stream corridor 1. Within this marginal zone this till unit ranges in thickness from 9-30m. Such thicknesses are not consistent with those recorded as subglacial traction deposits (Evans et al., 2006) but have been reconciled with the sub-marginal accretion of deforming sediment units (Boulton 1996a, b; Eyles et al., 2011; Evans et al., 2008). Furthermore within the northernmost limits of the study area till is considerably thinner and in some areas of the Turtlelake Upland bedrock is exposed. It is proposed, due the presence of a boulder pavement at the base of this unit, that the unit is analogous to a retreat phase till (where no advance phase till is preserved), consistent with the erosional zone of Boulton's (1996a, b) model. This boulder pavement can be traced to the southern margin of the SWSS and in this area is recorded within rather than at the contact of DU-6.

The characteristics of DU-6 thus lend support to Boulton's (1996a, b) model of till deposition patterns in former ice sheets. The thickening of DU-6 in association with terminal moraine ridges in the south of the study area suggests that: **1.** the snout was stable over a period of time that was sufficient during the Late Wisconsinan for incremental stacking of till to occur; and **2.** deformation of till was, at least within the marginal zone of this ice stream, significant in the net down glacier transfer of subglacial sediment. The characteristics of DU-6 thus provides the first evidence within southwestern Saskatchewan of large scale patterns of marginal till thickening. This evidence is consistent with site specific evidence of marginal ice stream thickening of glaciogenic sediments in Alberta (Evans et al., 2008) associated with the HPIS and CAIS. Distinct from both of these ice streams, till associated with Ice Streams 2A, B and C (DU-7) shows limited thickening within the study side. However given the limited extent of this till this is unsurprising. Interestingly when this surficial till (DU-7) is tracked southeast to the James Lobe it is reported to thicken considerably (Hallberg and Kemmus, 1986). Within eastern North Dakota (south corner of the James Lobe) the surficial unit associated with this ice stream reaches a thickness of up to 15m (Hallberg and Kemmus, 1986). It is therefore tentatively suggested that this ice stream system might also exhibit ice marginal thickening. Thus collectively this region of the LIS carries evidence of ice marginal thickening consistent with theoretical models.

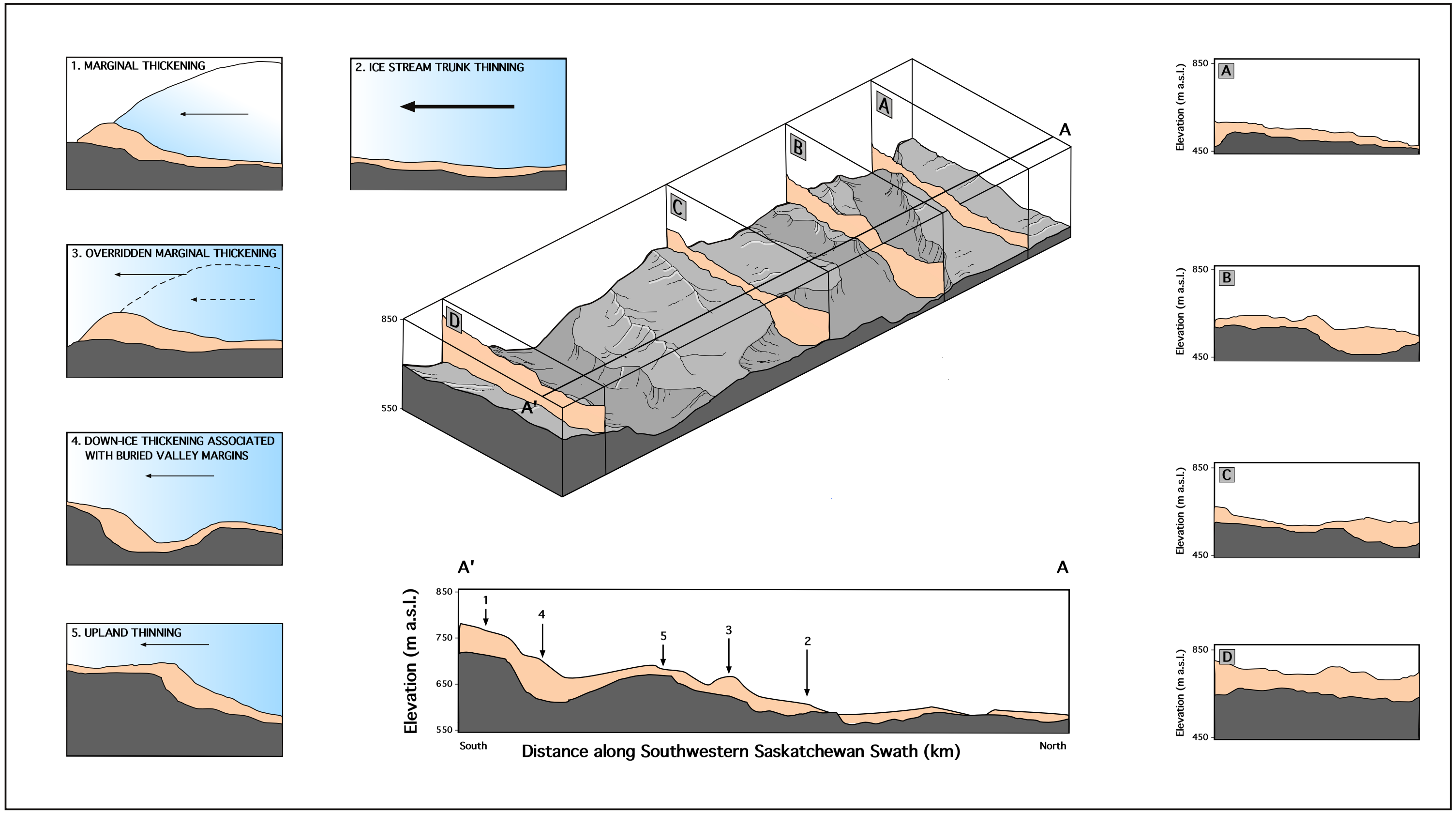


Figure 7.4: *Model of the proposed till deposition associated with fast flow. 3D and transect view of net sediment deposition across the SWSS (numbers on transect correspond to processes on the left). Processes responsible for variation in sediment thickness associated with fast ice flow:* ***1.*** *marginal thickening* ***2.*** *ice stream trunk thinning* ***3.*** *thickening as a result of overridden glacial-marginal landforms* ***4.*** *down-ice thickening associated with buried valley margins* ***5.*** *upland thinning.* ***A.*** *Low net till accumulation associated with the onset zone of an ice stream (Ice Stream 1) this region displays thin net sediment accumulation resulting from limited deposition and high level of erosion.* ***B.*** *Increased till deposition associated with marginal thickening of Ice Stream 1, additionally within this transect till deposition associated an overridden glacial margins is also recorded.* ***C.*** *Increased till deposition associated with down-ice side valley thickening is displayed on the east of this cross section, on the west of this cross section sediment deposition associated with upland thinning and upland thinning is displayed.* ***D.*** *High net till deposition associated with marginal thinning of till at the terminus of an ice stream (Ice Stream 1). While this model is based upon the deposition of sediment within the SWSS the process it represents may be applied in a more general sense to other palaeo-ice streams.*

The thickness of DU-6 at its southernmost limit is in considerable contrast to its thickness at the centre of the trunk zone of Ice Stream 1. For example north of Ice Stream 2B only a thin (average thickness 2.5 m) layer of DU-6 is present. The occurrence of thin tills at the centres of the fast-flow corridors, in many places unconformably overlying stratified sediments (Piotrowski et al., 2004), suggests that widespread till deformation may have been, in some locations within the SWSS, secondary to basal sliding in driving palaeo-ice streams. Nevertheless, the general thickening of tills towards the ice stream/lobe margins is consistent with the theory of subglacial deformation presented by Boulton (1996a, b).

In addition to these surficial tills several stratigraphically older till units (DU-4 and 5) also record a generalised pattern of thickening towards the south/southwest with thinning in the centre of Ice Stream 1. This characteristic of stratigraphically older tills is pertinent for two reasons. Firstly, if we employ DU-6 as an analogue and assume that wedge like thickening is the result of marginal thickening consistent with subglacial deformation theory, then these wedges suggest that during several time periods ice reached the southern border of Saskatchewan and may have occupied a similar flow path as that recorded by the surficial geology. Secondly, recognition of these characteristics within older units suggests that it may be a valuable indicator that has the potential for recognition of ice stream activity and dynamics from its regional stratigraphic architecture alone.

7.2.2 Localised thickening of subglacial sediments

In addition to a generalised thickening of subglacial sediment towards ice stream margins, more localised thickening not predicted within Boulton's (1996a, b) model, is recognised. This variation in sediment thickness is attributed to three processes: **1.** down-ice thickening associated with buried valley margins; **2.** upland thinning; and **3.** thickening as a result of overridden glacial-marginal landforms.

7.2.2.1 Down-ice thickening associated with buried valley margins

Substantial thickening of tills and plucked bedrock immediately down-ice of buried valleys also occurs within the SWSS. This implies that pre-existing sediment sequences (valley fills) are crucial to the subglacial generation of sediment. This is partially applicable to DU-6 and 5 that show increases of 12 m and 9 m respectively of till thickening in such situations.

It is likely that as ice advanced against the regional slope this would have resulted in proglacial lake formation and infilling as recorded by thick glaciolacustrine deposits (Proudfoot, 1985; Evans and Campbell, 1995) (and this study: see Section 6.2.11). Mass flow diamicton associated with the advection of sub-marginal till would have also accumulated over valleys. Thus it is possible that sand and gravels in these preglacial sediment sequences in the buried valleys acted as aquifers, locally evacuating subglacial meltwater, increasing the basal shear stress and thereby causing till thickening either by tectonic stacking or by freeze-on and later melt-out (e.g. Christofferson and Tulaczyk, 2003). In addition this also explains how the leakage of subglacial meltwater through preglacial valley aquifers could conceivably have caused ice stream deceleration and possibly surging within Ice Stream corridors 2A, B and C. However, this thickened sediment would have acted as a blockage to streaming ice further up flow which would have thickened and eventually advanced, streamlining its surface. Therefore, the beds of fast flowing trunk ice can contain multiple imprints of glacier advance phases, and ice streaming can mould and redistribute pre-existing materials so that a mosaic of bed deformation and sliding is produced by localised changes in 'continuity' of subglacial materials (Evans et al., 2006).

7.2.2.2 Upland thinning

In association with valley thickening, upland thinning of till sequences is also observed immediately down-ice of buried valleys. This process is particularly well observed in relation to DU-6 within the Battleford and Tyner buried valley systems and DU-5 within the Battleford Valley system. It is proposed that this thinning is a negative feedback resulting from down-ice valley thickening. In upland areas down-ice of buried valleys ice would no longer have been subject to the increased basal shear stress that resulted from local draining of subglacial meltwater and concomitant ice deceleration. Thus it would have been able to re-accelerate and once again thicken. Such ice thickening would have facilitating sliding and prohibited the build up of thick tills (Evans et al., 2006; Piotrowski and Kraus, 1997). In combination with this, many upland areas lack sand rich intertill deposits and the glacier bed would have been moving over clay-rich till. The clay-rich nature of these tills would have allowed the build up of high pore water pressures leading to decreased basal shear stress and localised glacier decoupling (Kamb, 1998). This illustrates the potential lithological control on not only the subglacial hydrological system, but also the mechanism for glacier motion across its bed, leaving a distinctive signature in the till architecture.

7.2.2.3 Thickening as a result of overridden glacial-marginal landforms

Unrelated to buried pre-glacial valleys and/or upland areas, stratigraphically older tills also thicken. Based on the similarities of these cases of abnormally thick till to those recorded by Evans et al. (2008) in southern Alberta in relation to overridden ridges, it is proposed that these localised thickenings mark former overridden glacier margin landforms. These thick till sequences likely record ice-marginal moraine construction and subglacial till thickening during ice sheet advance. One example of this occurs within DU-3 on the southern side of the North Battleford Valley. However the proposal that this type of thickening is related to overridden ice margins is made especially clear where till thickening coincides with overridden ridge complexes. This is illustrated in relation to the large lobate complex of overridden transverse ridges interpreted as terminal moraines. This ridge complex is associated with glaciotectonised bedrock and tills (specifically DU-4, which is present only within the margins of this lobate ridge complex), which are stacked at the centre of large transverse ridges.

7.2.2.4 Synthesis: ice stream till deposition

While the aim of this thesis was to evaluate the significance of depositional patterns in the context of Boulton's (1996a, b) theoretical model of regional till architecture associated with fast flow and ice streaming, several characteristics of the till architecture within the SWSS distinguish it from this pre-existing model. Moreover this model is developed as a theoretical representation of regional deposition patterns and, as this study provides the first case study and synthesis of 3D till architecture associated with regional palaeo-ice stream activity, a new ice stream till deposition model is proposed (Fig 7.4). This model is a synthesis of observations made within the SWSS and observations made by previous authors within the surrounding ice streams (e.g. Patterson, 1997, 1998; Evans et al., 2008, 2012). The model provides a generalised view of the pattern of deposition resulting from fast flow over an unlithified sediment bed and thus, much like geomorphic landsystem models of palaeo-ice stream imprints (Hart, 1999; Stokes and Clark, 1999, 2001), can be used to infer the dynamic behaviour of other former ice sheets from their sedimentary imprint.

This model can be visualised as the evolution of 5 regions each present in Figure 7.4. Region 1 encompasses the marginal area of an ice stream; in this location the model depicts a generalised sub-marginal till thickening. Region 2 depicts a region in which erosion/sliding are dominant and associated with the ice stream's fast flowing trunk region. It should be noted that while in this region deformation may have been subordinate to sliding of the glacier bed, this does not mean that deformation was absent; indeed it is likely that locally this process would also have been an important process within these trunk region. Regions 3, 4 and 5 illustrate localised processes and are thus spatially variable within a single ice sheet. Thickening as a result of overridden glacial-marginal landforms is displayed as Region 3. Region 4 depicts down-ice side valley thickening; in such locations till is expected to thicken significantly as a result of ice stream deceleration. Linked to Region 4, Region 5 encompasses areas of upland thinning resulting from fast flow.

7.3 Significance and wider implications

The results, interpretation and discussion of this study have several implications for: **1.** stratigraphy modelling; **2.** our understanding of palaeo-ice stream sedimentary signatures; and **3.** the dynamics of the south western Interior Plains sector of the LIS during the Late Wisconsinan deglaciation.

Firstly, this research has provided the first 3D regional stratigraphic reconstruction in relation to palaeo-ice streams. This reconstruction has proved to be particularly insightful in interpreting the depositional history of the SWSS. This has allowed interpretations of preglacial, Pre-Late Wisconsinan and Late Wisconsinan landscape evolution to be made. Unlike studies that rely on sediment exposures of a limited number of boreholes, 3D modelling of borehole data allowed a regional picture of ice stream activity, enabling the spatial and temporal evolution of the landscape to be simultaneously assessed. Furthermore this methodology facilitated links between data that are undoubtedly connected to glacial processes (till architecture) and landform assemblages related to glacial dynamics, including ice streaming.

Secondly, stratigraphic investigation has demonstrated that the dynamic behaviour of former ice sheets can be recognised from their sediment imprint. Criteria for palaeo-ice stream identification associated with sedimentary characteristics have historically been less well known (Stokes and Clark, 2001). This study proposes that, in addition to Stokes and Clark's (1999) geomorphological criteria, the till architecture of an ice stream may also be diagnostic in the identification of palaeo-ice streams, and that sub-marginal till thickening, as theorised by Boulton (1996a, b), and ice stream trunk and upland till thinning, as well as preglacial valley thickening are also potentially characteristic of palaeo-ice streams operating over soft unconsolidated sediments. Implementation of these criteria thus may provide a more robust subglacial landscape analysis that will improve the way in which glacial stratigraphic records and geomorphological evidence are analysed and used to identify former ice streams.

This study also has wider implications in terms of the dynamic behaviour of the southwestern LIS during the Late Wisconsinan. High resolution geomorphological mapping and stratigraphic analysis, which have built on the reconnaissance style mapping completed in the region previously (Ross et al., 2009; Ó Cofaigh et al., 2010), confirms interpretations of the superimposition of ice stream corridors. This suggests that a dynamic switch in flow direction occurred in this region and subsequently the persistent fast flow of Ice Stream 1 gave way to the more transitory surging system of Ice Streams 2A, B and C, indicating the possibility of periodic spatial and temporal instability of the ice drainage network within this region of the LIS during the Late Wisconsinan.

7.4 Evaluation of methods and data sources: limitations and further research

This study has provided a comprehensive analysis of the spatial dynamics and regional sedimentary architecture within the SWSS, however the techniques through which conclusions were reached are not without limitations. Such limitations have been alluded to throughout but are now discussed in detail.

7.4.1 Geomorphic Mapping

Geomorphic investigation of the SWSS was completed using a combination of SRTM and Landsat ETM+ imagery. SRTM data provided an excellent source of regional scale mapping, allowing suites of landforms and large single features to be easily identified. The elevation data contained within this data set and the simplicity with which visualisation manipulations can be undertaken make SRTM data an excellent resource for large scale mapping without which many landforms would not have been recognised based purely on Landsat ETM+ imagery. This was particularly important in identifying large areas of smoothed terrain covered in megalineations. However, like mapping in southwestern Saskatchewan previously completed by Ó Cofaigh et al. (2010), many smaller features such as esker networks and the intricacies of minor transverse ridges could not be identified using the SRTM imagery due to resolution constraints. In some cases these features could be easily identified when SRTM imagery was used in conjunction with higher resolution Landsat ETM+ imagery. One example of this is the identification of landforms that demarcate Ice Stream 2C that were previously unrecognised. However in a number of cases this combined mapping technique still has limitations. For example as discussed in section 5.1.5, smaller meltwater channels could have existed which could not be resolved and/or differentiated from sub-aerial rivers (active, seasonal or dry). This is also exemplified when looking at hummocky terrain where the resolution of imagery was insufficient to make significant insight into its morphology. To resolve such problems future work should aim to undertake ground truthing in particularly problematic areas, such as around the former flow path of Ice Stream 2C, thereby providing more evidence with which to interpret and identify certain features.

7.4.2 Stratigraphic analysis

The use of stratigraphic data was pivotal to this study. This data has provided a detailed regional view of past glacial and interglacial depositional records from which the history of the region and the dynamics and processes of the LIS through time was inferred. While this data has been valuable, there are several limitations with relate to its application. Firstly, as highlighted in sections 5.2 some units displayed very similar lithologic and geophysical properties and so could often only be differentiated with confidence when separated by stratified units. To increase the confidence of stratigraphic interpretations future research should aim to sample several high-resolution cores within southwestern Saskatchewan and investigate their mineralogic and petrologic properties, such as the composition of the very coarse-sand fraction, clast lithology, carbonate content and till geochemisty. Such analysis would not only provide a greater level of confidence when differentiating stratigraphic units, but would also provide insight into subglacial transport distance and thus glacial flow patterns.

Another limitation related to sedimentological analysis within this study is the tentative nature of correlation of stratigraphic units with those of previous studies. As stated in section 7.1.4 such correlations are very provisional, especially in relation to the oldest tills (DU-1, 2, 3 and 4). Therefore the most appropriate extension to this work may be a continuous transect across Alberta, Saskatchewan and Manitoba to correlate these deposits. In addition to limitations relating directly to the stratigraphic units, it is also clear that there is a significant lack of dating control within the units identified within the study area. The construction of a chronostratigraphy is of great importance in order to understand the timescale over which deposition of stratigraphic units occurred and would provide a robust test of the inferred stratigraphic chronology suggested in previous work (Christiansen, 1971) and used in this study.

7.4.3 Stratigraphic analysis: the application of Rockworks 16™

This study has employed Rockworks16 ™. The application of this program within this study has allowed a previously unprecedented view of the glacial stratigraphy of southwestern Saskatchewan to be produced. It is thus necessary to critically access the use of this program in reconstructing, analysing and interpreting sub-surface stratigraphy from a palaeoglaciological perspective.

One of the key 2D outputs used within this study were computer-generated isopach maps. These were particularly useful in interpreting till thickening and were created (using an inverse modelling logarithm) and reviewed to identify anomalous data points. Such points were evaluated and in many cases reinterpreted to create more consistent isopach trends. Isopach maps were also filtered to remove extremely thin parts of units, and thicknesses <2m were set to zero. This was particularly effective to remove outliers that were well outside the main part of the unit. These manipulations were all completed within Rockworks, demonstrating the very adaptive nature of the program. However one challenge in creating isopach maps is that in some areas boreholes did not penetrate the entire thickness of a unit. This produced a thin false region of an isopach map. This was partially evident in lower till units DU-1 and 2. In such areas the model was edited and thicknesses were estimated from the surrounding boreholes.

Another problem occurring in the production of isopach maps was the incidence of unrealistic contour trends. Contours are created based on the thickness of stratigraphic units from borehole measurements; these values are then extrapolated to the surrounding areas in a circular fashion. One limitation of this occurs where very thick and/or thin till units are recorded in only one borehole in a region. These values are often over extrapolated, creating unrealistic isopach contours within sections of the study area. Furthermore in areas where tills are topographically constrained (i.e. within buried valleys) the circular interpolation of stratigraphic units takes no account of such features, and hence in such areas maps were edited. Therefore as stated in section 4.2.4 the isopach maps produced within this study should be viewed as estimates of sediment thickness.

One of the most important stages in constructing stratigraphic models in Rockworks is the production of grids. Grids are created by assigning values to grid nodes between boreholes that correspond to a stratigraphic class based on the relative proximity of each grid node to the surrounding holes. A strength of this 3D gridding process is that the interpolated data in the resulting 3D grid have the appearance of a stratigraphic unit, with an aspect ratio that emphasises the horizontal dimension over the vertical. Also the method preserves the local variability of the stratigraphy in each drill hole with no smoothing or averaging. Thus, where data are abundant local variability is incorporated. One limitation of this type of numerical interpolation is the sensitivity to the distribution of the data, where values from an isolated drill hole tend to extrapolate outward to fill an inordinate amount of the model area. The effect is particularly noticeable where a small number of deep drill holes are interspersed with shallower holes. This point is exemplified within the northwest of the study area. In this region few boreholes are present and as a result when grid models are created a high degree of extrapolation is present. This is exemplified by the surface bedrock topography grid (Fig 5.5) where the model has produced clear steps in the topography and a flat zone in the very northwest of the study area.

In summary while there are still some limitations associated with the application of Rockworks16™, this program is extremely useful in the delineation of stratigraphic units, and the mapping of their extent, thickness and 3D distribution. Thus the approach within this study has proved to be a successful methodology with which to approach sub-surface stratigraphic modelling where borehole data are relatively abundant. Furthermore, its success within an area composed of complex stratigraphies with numerous till units suggests that this approach would be highly applicable in a number of sedimentological settings as a tool to assess sedimentary architecture. In order to extend the results of this study region, it would also seem sensible, given the availability of borehole data that exists within Saskatchewan and Alberta to apply the methods used in this study to model till architecture across other areas of the Canadian Prairies. Of particular interest would be 3D modelling of the HPIS and CAIS, and the Sand River area 73L of Alberta. This would allow a quantification of the volume of subglacial till emplaced during ice stream activity through time.

7.5 Summary

This chapter has reconstructed the Quaternary stratigraphy, surficial geomorphology and depositional history of the SWSS. 17 glacial and nonglacial units indicate that ice advanced across the study area 7 times. The thick tills in this region record the erosion, transport and deposition of unlithified sediment by ice streams and provide the first regional scale evidence to allow the recognition of the activity and dynamic behaviour of a former ice sheet and its ice streams from its till architecture.

8. Conclusions

The aims of this study were to: **i.** reconstruct the depositional history of the SWSS; **ii.** determine the spatial dynamics and regional sedimentary architecture of palaeo-ice streams; and **iii.** evaluate the significance of depositional patterns in the context of theoretical models of regional till architecture (Boulton, 1996a, b). With reference to these aims the following conclusions are drawn:

1. Detailed mapping over a 57,400 km^2 area of southwestern Saskatchewan reaffirms the previous proposal of a southwest trending ice stream demarcated by a corridor of megaflutes and MSGLs (Ice Stream 1) extending from the Canadian Shield to southwestern Saskatchewan. This corridor is cross cut by three (one previously unrecognised) south to southwest trending ice streams (Ice streams 2A, B and C).
2. The superimposition of these corridors suggests that a dynamic switch in flow direction occurred in this region and subsequently the persistent fast flow of Ice Stream 1 gave way to the more transitory surging system of Ice Streams 2A, B and C. This suggests the possibility of periodic instability within this region of the LIS during the Late Wisconsinan.
3. Analysis of the lithologic and geophysical characteristics of 197 borehole samples within these corridors revealed a superimposed glaciogenic sequence of 17 stratigraphic units.
4. A 3D stratigraphic model of the 57,400 km^2 swath was then constructed by extrapolating data away from boreholes using a nearest-neighbour approach. Using this model the thickness, extent and distribution of these stratigraphic units was delineated, allowing the depositional history of the region to be reconstructed and thus the extent of till emplaced during ice stream operation through time and space to be inferred.
5. Reconciling this regional till architecture with the regional geomorphology revealed that surficial units are spatially consistent with a dynamic switch in ice stream flow direction.
6. Thin tills at the centre of the trunk zone of Ice Stream 1 in many places lie unconformably over stratified sediments, suggesting widespread basal sliding may have been subordinate to deformation, but the general thickening of tills

towards the lobate terminal margins is consistent with subglacial deformation theory (Boulton 1996a, b).

7. Additionally variation in till thickness is also recognised on a more localised scale. These variations are attributed to three processes: **1.** down-ice side thickening associated with buried valley margins; **2.** upland thinning; **3.** thickening as a result of overridden glacial-marginal landforms.
8. The significance of newly interpreted patterns of ice stream deposition were then considered and a model of ice stream till deposition presented. This model provides a generalised view of the pattern of deposition resulting from fast flow over an unlithified sediment bed, which may be used to infer the dynamic behaviour of other former ice sheets from their stratigraphic imprint.

Overall this study has provided an in-depth stratigraphic and geomorphological assessment of the SWSS. The importance of integrating high resolution geomorphological mapping and regional till architectural evidence when analysing former terrestrial ice stream dynamics was deemed of critical importance. Thus future studies should consider combining these approaches when undertaking subglacial landscape analysis in order to improve understanding of the spatial and temporal dynamics of palaeo-ice streams.

References:

Adams, R. S. (2009) *An Integrated Approach for Paleo-Ice Stream Determination in Mid Continental Prairies, Saskatchewan, Canada*. Department of Earth Sciences. Masters Thesis. University of Waterloo.

Aber, J. S. (1993) Geomorphic and structural genesis of the Dirt Hills and Cactus Hills, southern Saskatchewan. In Aber, J. S. (ed.) (1993): *Glacio-tectonics and Mapping Glacial Deposits*. Canadian Plains Research Centre, Regina.

Alden, W. M. C. and Stebinger, E. (1913) Pre-Wisconsinan glacial drift in the region of glacier national park, Montana. *Bulletin of the Geological Society of America.* 24: 529–572.

Alley, R. D. (1989) Water-pressure coupling of sliding and bed deformation: I. Water system. *Journal of Glaciology*. 35: 108-118.

Alley, R. B. (1991) Deforming bed origin for southern Laurentide till sheets? *Journal of Glaciology*. 37: 67-76.

Alley, R. B. and Bindschadler, R. A. (2001) The West Antarctic ice sheet and sea-level change. In: Alley, R. B. and Bindschadler, R.A. (eds) (2001) The West Antarctic Ice Sheet: Behaviour and Environment. *American Geophysical Union, Antarctic Research Series.* 7: 1-11.

Alley, R. B., Blankenship, D. D., Bentley, C. R. and Rooney, S. T. (1986) Deformation of till beneath Ice Stream B, West Antarctica. *Nature*. 322: 57-59.

Alley, R. B., Cuffey, K. M., Everson, E. B., Strasser, J. C., Lawson, D. E. and Larson, J. C. (1997) How glaciers entrain and transport basal sediment: physical constraints. *Quaternary Science Reviews.* 16: 1017-1038.

Anandakrishnan, S., Blankenship, D. D., Alley, R. B. and Stoffa, P. L. (1998) Influence of subglacial geology on the position of a West Antarctic ice stream from seismic observations. *Nature.* 394: 62-65.

Anderson, J. B. and Oakes Fretwell, L. (2008) Geomorphology of the onset area of a paleo-ice stream, Marguerite Bay, Antarctic Peninsula. *Earth Surface Processes and Landforms*. 33: 503-512.

Andriashek, L. D. and Fenton, M. M. (1989) *Quaternary stratigraphy and surficial geology of the Sand River area 73L*. Alberta Research Council, Edmonton.

Atkinson, N., Utting, D. J. and Pawley, S. P. (2014) Glacial Landforms of Alberta. Alberta Geological Survey, *AER/AGS.* Map 604.

Barendregt, R. W., Irving, E., Christiansen, E. A., Sauer, E. K. and Schreiner, B. T. (1998) Stratigraphy and paleomagnetism of Late Pliocene and Pleistocene sediments in the

Wellsch Valley and Swift Current Creek areas, southwestern Saskatchewan, Canada. *Canadian Journal of Earth Sciences*. 35: 1347–1361.

Bell, R. (1884) Report on part of the basin of the Athabasca River, North-West Territory. *Geological Survey of Canada*. 35: 1882–1883.

Bell, R. (2008) The role of subglacial water in ice-sheet mass balance. *Nature Geoscience*. 1: 297-304.

Benn, D. I. and Evans, D. J. A. (2004) *A Practical Guide to the Study of Glacial Sediments.* Arnold, London.

Benn, D. I. and Evans, D. J. A. (2006) Subglacial megafloods: outrageous hypothesis or just outrageous? In: Knight, P. G. (ed.) (2006) *Glacier Science and Environmental Change.* Blackwell, London.

Bennett, M. R. (2003) Ice streams as the arteries of an ice sheet: their mechanics, stability and significance. *Earth-Science Reviews*. 61: 309-339.

Bennett, M.R., Huddart, D. and McCormick, T. (2000) An integrated approach to the study of glaciolacustrine landforms and sediments: a case study from Hagavatn, Iceland. *Quaternary Science Reviews*. 19: 633-665.

Bentley, C. R. (1987) Antarctic ice streams: a review. *Journal of Geophysical Research*. 92: 8843-8858.

Betcher, R. N., Matile, G., and Keller, G. (2005) Yes Virginia, there are buried valley aquifers in Manitoba. In: *Proceedings of the 58th Canadian Geotechnical Conference.* September 18–21, 2005, Saskatoon, Saskatchewan, Canada.

Bik, M. J. J. (1969) The origin and age of prairie mounds of southern Alberta. *Biuletyn Peryglacjalny*. 19: 85-130.

Bindschadler, R. A., Stephenson, S. N., MacAyeal, D. R. and Shabataie, S. (1987) Ice dynamics at the Mouth of Ice Stream B, Antarctica. *Journal of Geophysical Research.* 92: 8885-8894.

Blankenship, D. D., Bentley, C. R., Rooney, S. T. and Alley, R. B. (1986) Seismic measurements reveal a saturated porous layer beneath an active Antarctic ice stream. *Nature*. 322: 54-57.

Boulton, G. S. (1972) Modern arctic glaciers as depositional models for former ice sheets. *Journal of Geological Society of London*. 128: 361-93.

Boutlon, G. S. (1987) A theory of drumlin formation by subglacial sediment deformation. In: Menzies, J and Rose, J. (eds) (1987) *Drumlin symposium*. Balkema, Rotterdam.

Boulton, G. S. (1996a) Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation. *Journal of Glaciology.* 42: 43-62.

Boulton, G.S. (1996b) The origin of till sequences by subglacial sediment deformation beneath mid-latitude ice sheets. *Annals of Glaciology.* 22: 75-84

Boulton, G. S. and Clark, C. D. (1990) A highly mobile Laurentide Ice Sheet revealed by satellite images of glacial lineations. *Nature*. 346: 813–817.

Boulton, G. S. and Hindmarsh, R. C. A. (1987) Sediment Deformation Beneath Glaciers: Rheology and Geological Consequences. *Journal of Geophysical Research.* 92: 9059-9082.

Boulton, G. S. and Jones, A. S. (1979) Stability of temperate ice sheets resting on beds of deformable sediment. *Journal of Glaciology*. 24: 29-43.

Boulton, G. S., Smith G. D., Jones, A. S. and Newsome, J. (1985) Glacial geology and glaciology of the last mid-latitude ice sheets. *Journal of the Geological Society*. 142: 447-474.

Boulton, G. S., Dobbie, K. E. and Zatsepin, S. (2001a) Sediment deformation beneath glaciers and its coupling to the subglacial hydraulic system. *Quaternary International.* 86: 3-28.

Boulton, G. S., Dongelmans, P., Punkari, M. and Broadgate, M. (2001b) Palaeoglaciology of an ice sheet through a glacial cycle: the European ice sheet through the Weichselian. *Quaternary Science Reviews*. 20: 591-625.

Briner, J. P. (2007) Supporting evidence for the New york drumlin field that elongate subglacial bedforms indicate fast ice flow. *Boreas*. 36: 143-147.

Calhoun, R. H. H. (1906) Montana lobe of the Keewatin ice sheet. *United States Geological Survey*. 50: 62-68.

Canals, M., Urgeles, R. and Calafat, A. M. (2000) Deep sea-floor evidence of past ice streams off the Antarctic Peninsula. *Geology*. 28: 31-34.

Campbell, J. C. (1986a) Surficial Geology of the Kindersley Area (72N). *Saskatchewan: Saskatchewan Research Council, Sedimentary Resources.* Map 1:250,000 scale.

Campbell, J. C. (1986b) Surficial Geology of the Melville Area (72K). *Saskatchewan: Saskatchewan Research Council, Sedimentary Resources.* Map 1:250,000 scale.

Campbell. J. C. (1987a) Surficial Geology of the St. Walburg Area (73F). *Saskatchewan: Saskatchewan Research Council, Sedimentary Resources.* Map 1:250,000 scale.

Campbell, J. C. (1987b) Surficial Geology of the North Battleford Area (73C). *Saskatchewan: Saskatchewan Research Council, Sedimentary Resources.* Map 1:250,000 scale.

Christiansen, E. A. (1968a) A thin till in west-central Saskatchewan, Canada. *Canadian Journal of Earth Science.* 5: 329–336.

Christiansen, E. A. (1968b) Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada. *Canadian Journal of Earth Sciences.* 5: 1167–1173.

Christiansen, E. A. (1971) Tills in Southern Saskatchewan, Canada. In: Goldthwait, R. P. (ed.) (2004*) Till: A Symposium*. Ohio State University Press, Columbus.

Christiansen, E. A. (1972) *Cross section of drift and bedrock between Condie and Craven, Saskatchewan.* Saskatchewan Research Council. Geology Division, Cross Section 1.

Christiansen, E. A. (1979) The Wisconsinan deglaciation of southern Saskatchewan and adjacent areas. *Canadian Journal of Earth Sciences*. 16: 913-938.

Christiansen, E. A. (1987) Verendrye valley and the Glidden esker, Saskatchewan: subglacial and ice walled features in southwestern Saskatchewan, Canada. *Canadian Journal of Earth Sciences.* 24. 170-176.

Christiansen, E. A. (1992) Pleistocene stratigraphy of the Saskatoon area, Saskatchewan, Canada – an update. *Canadian Journal of Earth Sciences*. 29: 1767–1778.

Christoffersen, P. and Tulaczyk, S. (2003) Thermodynamics of basal freeze-on: predicting basal and subglacial signatures of stopped ice streams and interstream ridges. *Annals of Glaciology*. 36: 233-243.

Clark, C. D. (1993) Mega-scale glacial lineations and cross-cutting ice-flow landforms. *Earth Surface Processes and Landforms*. 18: 1-29.

Clark, C. D. (1994) Large-scale ice-moulding: a discussion of genesis and glaciological significance. *Sedimentary Geology*. 91: 253-268.

Clark, C. D. (1997) Reconstructing the evolutionary dynamics of former ice sheets using multi-temporal evidence, remote sensing and GIS. *Quaternary Science Reviews*. 16: 1067-1092.

Clark, C. D. and Stokes, C. R. (2003) Palaeo-ice stream landsystem. In: Evans, D. J. A. (ed.) *Glacial Landsystems*. Arnold, London.

Clark, P. U. (1991) Striated clast pavements: products of deforming subglacial sediment? *Geology*. 19: 530-533.

Clark, P. U. (1992) Surface form of the southern Laurentide Ice Sheet and its implications to ice-sheet dynamics. *The Geological Society of American Bulletin*. 104: 595-605.

Clayton, L. (1967) Stagnant glacier features of the Missouri Coteau in North Dakota. North Dakota Geological Survey. *Miscellaneous Series*. 30: 25-46.

Clayton, L. and Moran, S. R. (1982) Chronology of Late Wisconsinan glaciation in middle North America. *Quaternary Science Reviews.* 1: 55–82.

Clayton, L., Teller, J. T. and Attig, J. W. (1985) Surging of the south- western part of the Laurentide Ice Sheet. *Boreas.* 14: 235–241.

Conway, H., Catannia, G., Raymond, C. F., Gades, A. M., Scambos, T. A., and Engelhardt, H. (2002) Switch of flow direction in an Antarctic ice stream. *Nature*. 419: 465-467.

Cummings, D. I., Russell, H. A. J. and Sharpe, D. R. (2012) Buried valleys and till in the Canadian Prairies: geology, hydrogeology, and origin. *Geological Survey of Canada*. 22: 2012-2014.

De Angelis, H. and Kleman, J. (2005) Palaeo-ice streams in the northern Keewatin sector of the Laurentide ice sheet. *Annals of Glaciology*. 42: 135-144.

De Angelis, H. and Kleman, J. (2007) Palaeo-ice streams in the Foxe/Baffin sector of the Laurentide Ice Sheet. *Quaternary Science Reviews*. 26: 1313-1331.

De Angelis, H. and Kleman, J. (2008) Palaeo-ice stream onsets: examples from the north- eastern Laurentide Ice Sheet. *Earth Surface Processes and Landforms*. 33: 560-572.

Denton, G. H. and Hughes, T. J. (1981) The Arctic Ice Sheet: An outrageous hypothesis. In: Denton, G. H. and Hughes, T. J. (eds.) (1981) The Last Great Ice Sheets. Wiley, New York.

Dowdeswell, J. A., Ottesen, D. and Rise, L. (2006) Flow switching and large-scale deposition by ice streams draining former ice sheets. *Geology*. 34: 313-316.

Dowdeswell, J. A., O´Cofaigh, C. and Pudsey, C. J. (2004) Thickness and extent of the subglacial till layer beneath an Antarctic palaeo-ice stream. *Geology*. 32: 13–16.

Dyke, A. S. and Prest, V. K. (1987) Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. *Geographie physique et Quaternaire*. 41: 237–263.

Dyke, A. S. and Morris, T. F. (1988) Drumlin fields, dispersal trains, and ice streams in Arctic Canada. *Canadian Geographer*. 32: 86-90.

Dyke, A. S., Morris, T. F., Green, D. E. C. and England, J. H. (1992) Quaternary geology of Prince of Wales Island, central Canadian Arctic. *Geological Survey of Canada*. 2: 433-438.

Dyke, A. S., Andrews, J. T., Clark, P. U., England, J., Miller, G. H., Shaw, J. and Veillette. J. J. (2002) The Laurentide and Innuitian Ice Sheets during the Last Glacial Maximum. *Quaternary Science Reviews*. 21: 9-31.

Dyke, A. S., Moore, A. and Robertson, L. (2003) Deglaciation of North America. Geological Survey of Canada Open File 1574. Digital data available at: *http://geopub.nrcan.gc.ca.*

Echelmeyer, K. A., Harrison, W. D., Larsen, C. and Mitchell, J. E. (1994) The role of the margins in the dynamics of an active ice stream. *Journal of Glaciology*. 40: 527-538.

Englehardt, H. and Kamb, B. (1998) Basal sliding of Ice Stream B, West Antarctica. *Journal of Glaciology*. 44: 223-230.

Engelhardt, H., Humphrey, N., Kamb, B. and Fahnestock, M. (1990) Physical Conditions at the Base of a Fast Moving Antarctic Ice Stream. *Science*. 248: 57-59.

Evans, D. J. A. (2003) Ice-marginal terrestrial landsystems: active temperate glacier margins. In: Evans, D. J. A. (ed.) (2003) *Glacial Landsystems*. Arnold, London.

Evans, D. J. A and Campbell, I. A. (1995) Quaternary stratigraphy of the buried valleys of the lower Red Deer River, Alberta, Canada. *Journal of Quaternary Science*. 10: 123-148.

Evans, D. J. A and Rea, B. R. (1999) Geomorphology and sedimentology of surging glaciers: a landsystems approach. *Annals of Glaciology*. 28: 75-82.

Evans, D. J. A. and Rea, B. R. (2003) Surging glacier landsystem. In: Evans, D. J. A. (ed.) (2003) *Glacial Landsystems*. Arnold, London.

Evans. D. J. A. and Twigg, D. R. (2002) The active temperate glacial landsystem: a model based on Breidamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews*. 21: 2143-2177.

Evans, D. J. A., Philips, E. R., Hiemstra, J. F. and Auton, C. A. (2006) Subglacial till: formation, sedimentary characteristics and classification. *Earth Science Reviews*. 78: 115-176.

Evans, D. J. A., Hiemstra, J. F. and O'Cofaigh. C. (2007) An assessment of clast macrofabrics in glacigenic sediments based on A/B plane data. *Geografiska Annaler Series a-Physical Geography*. 89: 103-120.

Evans, D. J. A., Clark, C. D. and Rea, B. R. (2008) Landform and sediment imprints of fast glacier flow in the southwest Laurentide Ice Sheet. *Journal of Quaternary Science.* 23: 249–272.

Evans, D. J. A., Hiemstra, J. F., Boston, C. M., Leighton, I., Ó Cofaigh, C. and Rea, B. R. (2012) Till stratigraphy and sedimentology at the margins of terrestrially terminating ice streams: case study of the western Canadian prairies and high plains. *Quaternary Science Reviews.* 46: 80–125.

Evans, D. J. A., Young, N. J. P. and Ó Cofaigh, C. (2014) Glacial geomorphology of terrestrial-terminating fast flow lobes/ice stream margins in the southwest Laurentide Ice Sheet. *Geomorphology*. 204: 86-113.

Evans, D. J. A., Ewertowski, M. and Orton, C. (2015) Fláajökull (north lobe), Iceland: active temperate piedmont lobe glacial landsystem. *Journal of Maps*. 11:415-431

Evans, D. J. A. Storrar, R. D. and Rea, B. (2016) Crevasse-squeeze ridge corridors: diagnostic features of late-stage palaeo-ice stream activity. *Geomorphology*. 258: 40-50.

Everest, J., Bradwell, T. and Golledge, N. (2005) Subglacial Landforms of the Tweed Palaeo-Ice Stream. *Scottish Geographical Journal*. 121: 163-173.

Eyles, N., Boyce, J. I., Barendregt, R. W. (1999) Hummocky moraine: sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology.* 123: 163-174.

Eyles, N., Eyles, C., Menzies, J. and Boyce, J. (2011) End moraine construction by incremental till deposition below the Laurentide Ice Sheet: Southern Ontario, Canada. *Boreas*. 40: 92-104.

Fenton, M. M. (1984) Quaternary stratigraphy of the Canadian prairies. In:. R.J. Fulton (ed.) (1984) Quaternary Stratigraphy of Canada - A Canadian Contribution to IGCP Project 24. *Geological Survey of Canada*. 2: 57-68.

Fisher, D. A., Reeh, N. and Langley, K. (1985) Objective reconstructions of the Late Wisconsinan Laurentide Ice Sheet and the significance of deformable beds. Geographie physique et Quaternaire. 39: 229–238.

Foster, J. H. and Stalker, A. Mac, S. (1976) Paleomagnetic stratigraphy of the Wellsch Valley site, Saskatchewan. *Geological Survey of Canada*. 1: 98-107.

Fulton, R. J., Fenton, M. M. and Rutter, N. W. (1986) Summary of Quaternary Stratigraphy and Histroy, Western Canada. *Quaternary Science Reviews.* 5: 229-241.

Golledge, N. R. and Stoker, M. S. (2006) A palaeo-ice stream of the British Ice Sheet in eastern Scotland. *Boreas*. 35: 1-13.

Grant, N. M. (1997) *Genesis of the North Battleford Fluting Field, West-Central Saskatchewan.* Department of Earth and Atmospheric Sciences. Masters Thesis. University of Alberta.

Gravenor, C. P. and Meneley, W. A. (1958) Glacial flutings in central and northern Alberta. American. *Journal of Science*. 256: 715–728.

Gravenor, C. P. and Kupsch, W. 0. (1959) Ice-disintegration features in western Canada. *Journal of Geology*. 67: 48-64.

Hallberg, G. R. and Kemmis, T. J. (1986) Stratigraphy and correlation of the glacial deposits of the Des Moines and James lobes and adjacent areas in North Dakota, South Dakota, Minnesota, and Iowa. *Quaternary Science Reviews*. 5: 65–68.

Hambrey, M. J. and Dowdeswell, J. A. (1994) Flow regime of the Lambert Glacier-Amery Ice Shelf system, Antarctica: structural evidence from Landsat imagery. *Annals of Glaciology*. 20: 401-406.

Hart, J. K. (1999) Identifying fast ice flow from landform assemblages in the geological record: a discussion. *Annals of Glaciology*. 28: 59-66.

Hindmarsh, R. C. A. and Stokes, C. R. (2008) Formation mechanisms for ice-stream lateral shear margin moraines. *Earth Surface Processes and Landforms*. 33: 610- 626.

Hodge, S. M. and Doppelhammer, S. K. (1996) Satellite imagery of the onset of streaming flow of ice streams C and C, West Antarctica. *Journal of Geophysical Research*. 101: 6669-6677.

Hodgson, D. A. (1994) Episodic ice streams and ice shelves during retreat of the northwestern most sector of the Late Weichselian Laurentide ice sheet over the central Canadian Arctic Archipelago. *Boreas*. 23: 14-28.

Hughes, T. J. (1977) West Antarctic ice streams. *Review of Geophysics and Space Physics*. 15: 1-46.

Hulbe, C. L. and Fahnestock, M. A. (2004) West Antarctic ice stream discharge variability: Mechanism, controls and pattern of grounding line retreat. *Journal of Glaciology*. 50: 471-484.

Jacobel, R. W., Scambos, T., Raymond, C. F. and Gades, A. M. (1996) Changes in the configuration of ice stream flow from the West Antarctic Ice Sheet. *Journal of Geophysical Research*. 101: 5499-5504.

Jansson, K. N., Stroeven, A. P. and Kleman, J. (2003) 'Configuration and timing of Ungava Bay ice streams, Labrador-Ungava, Canada', Boreas, 32: 256-262

Jarvis, A., Reuter, H.I., Nelson, A. and Guevara, E. (2008) *Hole-filled seamless SRTM data V4.* International Centre for Tropical Agriculture (CIAT).

Jennings, C. E. (2006) Terrestrial ice stream – a view from the lobe. *Geomorphology*. 75: 100-124.

Joughin, I., Gray, L., Bindschadler, R., Price, S., Morse, D., Hulbe, C., Mattar, K. and Werner, C. (1999) Tributaries of West Antarctic Ice Streams Revealed by RADARSAT Interferometry. *Science*. 150: 283-286.

Joughin, I., Fahnestock, M., MacAyeal, D., Bamber, J. and Gogineni, P. (2001) Observation and analysis of ice flow in the largest Greenland ice stream. *Journal of Geophyscial Research.* 106: 34021-34034.

Kamb, B. (2001) Basal zone of the West Antarctic ice streams and its role in lubrication of their rapid motion. In: Alley, R. D. and Binschadler, R. (eds.) (2001) *The West Antarctic Ice Sheet: Behaviour and Environment*. Antarctic Research Series, American Geophysical Union, Washington.

King, E. C., Hindmarsh, R. C. A. and Stokes, C. R. (2009) Formation of mega-scale glacial lineations observed beneath a West Antarctic ice stream. *Nature Geoscience.* 2: 585-588.

Klassen, R. W. (1979) Pleistocene geology and geomorphology of the Riding Mountain and Duck Mountain areas, Manitoba-Saskatchewan. *Geological Survey of Canada. Memoir 396.*

Klassen, R. W. (1989) Quaternary geology of the southern Canadian interior plains. In: Fulton, R. J. (ed.) (1989) *Quaternary Geology of Canada and Greenland*. Geological Society of America, Colorado.

Klassen, R. W. (1991) Surficial geology and drift thickness, Cypress Lake (72F) Saskatchewan*. Geological Survey of Canada*. Map 1766A (1:250,000).

Klassen, R. W. (1992) Nature, origin and age relationships of landscape complexes in southwestern Saskatchewan. *Geographie Physique Et Quaternaire*. 46: 361-388.

Klassen, R. W. (1994) Late Wisconsinan and Holocene History of southwestern Saskatchewan. *Canadian Journal of Earth Sciences*. 31: 1822-1837.

Kleman, J. and Borgström, I. (1996) Reconstruction of paleao-ice sheets: the use of geomorphological data. *Earth Surface Processes and Landforms*. 21: 893-909.

Kleman, J. and Glasser, N. F. (2007) The subglacial thermal organisation (STO) of ice sheets. *Quaternary Science Reviews.* 26: 585-597.

Krüger, J. (1993) Moraine-ridge formation along a stationary ice front in Iceland. Boreas. 22: 101-9.

Krüger, J. (1994) Glacial processes, sediments, landforms, and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica*. 21: 1-233.

Kulig, J. J. (1996) The glaciations of the Cypress Hills of Alberta and Saskatchewan and its regional implications. *Quaternary International*. 32: 53–77.

Kupsch, W. O. (1962) Ice thrust ridges in western Canada. *Journal of Geology*. 70: 582-588.

Kuvaas, B. and Kristoffersen, Y. (1991) The Crary Fan: A trough-mouth fan on the Weddell Sea continental margin, Antarctica. *Marine Geology*. 97: 345-362.

Lillesand, T. M, R., Kiefer, W. and Chipman, J. W. (2008) *Remote Sensing and Image Interpretation*. New Jersey, John Wiley and Sons.

Livingstone, S J., Evans, D. J. A., Ó Cofaigh, C., Hopkins, J. (2010) The Brampton kame belt and Pennine escarpment meltwater channel system (Cumbria, UK): morphology, sedimentology and formation. *Proceedings of the Geologists' Association*. 121: 423-443.

Lowe, A. L. and Anderson, J. B. (2002) Reconstruction of the West Antarctic Ice Sheet in Pine Island Bay during the last Glacial Maximum and its subsequent retreat history. *Quaternary Science Reviews*. 21: 1879-1897.

Lüthi, M., Funk, M., Iken, A., Gogineni, S. and Truffer, M. (2002) Mechanisms of fast flow in Jakobshavn Isbrae, West Greenland. Part III. Measurements of ice deformation, temperature and cross-borehole conductivity in boreholes to the bedrock. *Journal of Glaciology*. 48: 369-385.

Maathuis, H. and Thorleifson, L.H. (2000) *Potential impact of climate change on Prairie groundwater supplies: review of current knowledge*. Saskatchewan Research Council, Publication. Saskatchewan.

Margold, M., Stokes, C. R. and Clark, C. D. (2015a) Ice streams in the Laurentide Ice Sheet: identification, characteristics and comparison to modern ice sheets. *Earth-Science Reviews*. 143: 117-146.

Margold, M., Stokes, C. R. and Kleman, J. (2015b) Ice streams in the Laurentide Ice Sheet: a new mapping inventory. *Journal of Maps*. 11: 380-395.

Marshall, S. J., Clarke, G. K. C., Dyke, A. S. and Fisher, D. A. (1996) Geologic and topographic controls on fast flow in the Laurentide and Cordilleran Ice Sheets. *Journal of Geophysical Research*. 101: 17827–17839.

Mathews, W. H. (1974) Surface Profiles of the Laurentide Ice Sheet in its Marginal Areas. *Journal of Glaciology*. 13: 37-43.

Mathews, J. A., Cornish, R. and Shakesby, R. A. (1979) 'Saw-tooth' moraines in front of Bodalsbreen, southern Norway. *Journal of Glaciology.* 22: 535-546.

McConnell, R. G. (1885) Report on the Cypress Hills, Wood Mountain, and adjacent country. *Geological Survey of Canada*. 1; 89-106.

Meneley, W. A. (1964) *Geology of the Melford area (73A) Saskatchewan*. Ph.D. thesis University of Illinois.

Miall, L. (2015) *Stratigraphy: A Modern Synthesis*. Springer, London.

Mosola, A. B. and Anderson, J. B. (2006) Expansion and rapid retreat of the West Antarctic Ice Sheet in eastern Ross Sea: possible consequence of over-extended ice streams? *Quaternary Science Reviews*. 25: 2177-2196.

Mossop, G. D. and Shetsen, I. (1994) Geological Atlas of the Western Canada Sedimentary Basin. Digital data available at: *http://ags.gov.ab publications/wcsb_atlas/atlas.ca.* (accessed: 23/05/2016).

NASA (2001) SRTM Topography. http://e)srp01u.ecs.nasa.gov/srtm/version2/Documentation/SRTM_Topo.pdf. (accessed: 23/05/2016).

Nichols, G. (2009) *Sedimentology and Stratigraphy*. Wiley-Blackwell, London.

Ó Cofaigh, C., Pudsey, C. J., Dowdeswell, J. A. and Morris, P. (2002) Evolution of subglacial bedforms along a palaeo-ice stream, Antarctic Peninsula continental shelf. *Geophysical Research Letters.* 29: 29-36.

Ó Cofaigh, C., Taylor, J., Dowdeswell, J. A. and Pudsey, J. A. (2003) Palaeo-ice streams, trough mouth fans and high-latitude continental slope sedimentation. *Boreas*. 32: 37-55.

Ó Cofaigh, C., Dowdeswell, J. A., Allen, C. S., Hiemstra, J. F., Pudsey, C. J., Evans, J. and Evans, D. J. A. (2005) Flow dynamics and till genesis associated with a marine- based Antarctic palaeo-ice stream. *Quaternary Science Reviews*. 24: 709-740.

Ó Cofaigh, C., Dowdeswell, J. A., Evans, J. and Larter, R. D. (2008) Geological constraints on Antarctic palaeo-ice-stream retreat. *Earth Surface Processes and Landforms.* 33: 513-525.

Ó Cofaigh, C., Evans, D. J. A. and Smith. R. (2010) Large-scale reorganization and sedimentation of terrestrial ice streams during late Wisconsinan Laurentide Ice Sheet deglaciation. *Geological Society of America.* 122: 743-756.

Ottesen, D., Stokes, C. R., Rise, L. and Olsen, L. (2008) Ice-sheet dynamics and ice streaming along the coastal parts of northern Norway. *Quaternary Science Reviews.* 27: 922-940.

Patterson, C. J. (1997) Southern Laurentide ice lobes were created by ice streams: Des Moines Lobe in Minnesota, USA. *Sedimentary Geology*. 111: 249-261.

Patterson, C. J. (1998) Laurentide glacial landscapes: The role of ice streams. *Geology*. 26: 643-646.

Paulson, Q. F. (1982) Guide to North Dakota's ground-water resources; United States Geological Survey. *Water-Supply Papers*. 2: 2236-2225.

Payne, A. J. and Dongelmans, P. W. (1997) Self-organisation in the thermomechanical flow of ice sheets. *Journal of Geophysical Research*. 102: 12219-12233.

Piotrowski, J. A., Doring, U., Harder, A., Qadirie, R., Wenghofer, S. (1997) 'Deforming bed conditions on the Danischer Wohld Peninsula, northern Germany': Comments. *Boreas.* 26: 73–77.

Piotrowski, J. A., Mickelson, D. M., Tulaczyk, S., Krzyszowski, D. and Junge, F. W. (2001) Were deforming beds beneath past ice sheets really widespread? *Quaternary International.* 86: 139-150.
Piotrowski, J. A., Larsen, N. J. and Junge, F. W. (2004) Reflections on soft subglacial beds as a mosaic of deforming and stable spots. *Quaternary Science Reviews*. 23. 993-1000.

Price, R. J. (1970) Moraines of Fjallsjoull, Iceland. *Arctic and Alpine Research*. 2: 27-42.
Price, S.F. and Whillans, I.M. (2001) Crevasse patterns at the onset to Ice Stream B, West Antarctica. *Journal of Glaciology*. 47: 29-36.

Proudfoot, D. N. (1985) *A lithostratigraphic and genetic study of Quaternary sediments in the vicinity of Medicine Hat, Alberta*. Ph.D. thesis. University of Alberta.

Rains, R. B., Kvill . D. and Shaw. J. (1999) Evidence and some implications of coalescent Cordilleran and Laurentide glacier systems in western Alberta. *In A World of Real Places: Essays in Honour of William C. Wonders.* University of Alberta.

Ramirez, E. (2006) Shuttle Radar Topography Mission: *The mission to map the world. Project Formulation and Management Applications (PFMA) Group*. http://www2.jpl.nasa.gov/srtm/. (accessed: 23/05/2016)

Raymond, C. F., Echelmeyer, K. A., Whillans, I. M., Doake, C. S. M. (2001) Ice stream shear margins. In: Alley, R. B., Bindschadler, R. A. (eds) (2001) *The West Antarctic Ice Sheet: Behaviour and Environment*. American Geophysical Union, Washington.

Retzlaff, R. and Bentley, C. R. (1993) Timing of stagnation of Ice Stream C, West Antarctica, from short-pulse radar studies of buried surface crevasses. *Journal of Glaciology*. 39: 553-561.

Rignot, E. (2006) Changes in ice dynamics and mass balance of the Antarctic ice sheet. *Philosophical Transactions of the Royal Society*. 364: 1637-1655.

Rignot, E., Mouginot, J. and Scheuchl, B. (2011) Ice flow of the Antarctic Ice Sheet. *Science*. 333: 1427-1430.

Roberts, D. H. and Long, A. J. (2005) Streamlined bedrock terrain and fast ice flow, Jakobshavn Isbrae, West Greenland: implications for ice stream and ice sheet dynamics. *Boreas*. 34: 25-42.

Rockware Inc., (2013) 3.23 User Manual. http://rockware.com/usermanual9.88.73 (accessed: 23/05/2016)

Rose, K. E. (1979) Characteristics of ice flow in Marie-Byrd Land, Antarctica. *Journal of Glaciology*. 24: 63-75.

Ross, M., Campbell, J., Parent, M. and Ross, R. (2009) Palaeo-ice streams and the subglacial landscape mosaic of the North American mid-continental prairies. *Boreas*. 38: 421-439.

Rutherford, R. L. (1937) Saskatchewan gravels and sands in central Alberta. *Transcrpts of the Royal Society of Canada*. 31: 81-95.

Sauer, E. K. and Christiansen, E. A. (1991) Preconsolidation pressures in the Battleford Formation, Southern Saskatchewan, Canada. *Canadian Journal of Earth Sciences*. 28: 1613–1623.

Sauer, E. K., Egeland, A. K. and Christiansen, E. A. (1993) Pre-consolidation of tills and intertill clays by glacial loading in southern Saskatchewan, Canada. *Canadian Journal of Earth Sciences.* 30: 420–433.

Saskatchewan Geological Survey (2003) Geology, and Mineral and Petroleum Resources of Saskatchewan. *Saskatchewan Industry and Resources.* Miscellaneous Report 2003-7.

Scott, J. S. (1976) Geology of Canadian tills. In: Leggett, R. F. (ed.) (1979) *Glacial Till, 50–68.* Royal Society of Canada, Ottawa.

Sharp, M. J. (1985) 'Crevasse-fill' ridges: a landform type characteristic of surging glaciers? *Geografiska Annaler*. 67: 213-220.

Shaw, J. (1983) Drumlin formation related to inverted meltwater erosional marks. *Journal of Glaciology*. 29: 461-479.

Shaw, J. (1989) Drumlins, subglacial meltwater floods, and ocean responses. *Geology*. 17: 853-856.

Shaw, J. (1994) Hairpin erosional marks, horseshoe vortices and subglacial erosion. *Sedimentary Geology*. 91: 269-283.

Shaw, J. and Sharpe, D. R. (1987) Drumlin formation by subglacial meltwater erosion. *Canadian Journal of Earth Sciences*. 24: 2316-2322.

Shaw, J., Pugin, A. and Young, R. R. (2008) A meltwater origin for Antarctic shelf bedforms with special attention to megalineations. *Geomorphology*. 102: 364-375.

Shipp, S. S., Anderson, J. B. and Domack, E. W. (1999) Late Pleistocene-Holocene retreat of the West Antarctic Ice Sheet system in the Ross Sea. Part 1; Geophysical Results. *Geological Society of America Bulletin*. 111: 1486-1516.

Smith, M. J., Clark, C. D. (2005) Methods for the visualization of digital elevation models for landform mapping. *Earth Surface Processes and Landforms*. 30: 885-900.

Stalker, A. M. (1960) Ice-pressed drift forms and associated deposits in Alberta. *Geological Survey of Canada Bulletin* 57: 30-38.

Stalker, A. M. (1961) Buried valleys in central and southern Alberta. *Geological Survey of Canada.* 15: 60-32.

Stalker, A. M. (1968) Identification of Saskatchewan Gravels and Sands. *Canadian Journal of Earth Sciences.* 5: 155-163.

Stalker, A. M. (1973) The large interdrift bedrock blocks of the Canadian Prairies. *Geological Survey Canada*. 75: 421-422.

Stokes, C. R. (2002) Identification and mapping of palaeo-ice stream geomorphology from satellite imagery: implications for ice stream functioning and ice sheet dynamics. *International Journal of Remote Sensing*. 23: 1557-1563.

Stokes, C. R. and Clark, C. D. (1999) Geomorphological criteria for identifying Pleistocene ice streams. *Annals of Glaciology.* 28: 67-74.

Stokes, C. R. and Clark, C. D. (2001) Palaeo-ice streams. *Quaternary Science Reviews.* 20: 1437–1457.

Stokes, C. R. and Clark, C. D. (2002a) Ice Stream Shear Margin Moraines', Earth *Surface Processes and Landforms*. 27: 547-558.

Stokes, C. R. and Clark, C. D. (2002b) Are long subglacial bedforms indicative of fast ice flow? *Boreas*. 31: 239-249

Stokes, C. R. and Clark, C. D. (2003) The Dubawnt Lake palaeo-ice stream: evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding the controls on ice-stream location and vigour. *Boreas*. 32: 263-279.

Stokes, C. R. and Clark, C. D. (2004) Evolution of late glacial ice-marginal lakes on the northwestern Canadian Shield and their influence on the location of the Dubawnt Lake palaeo-ice stream. *Palaeogeography, Palaeoclimatology and Palaeoecology.* 215: 155-171.

Stokes, C. R., Clark, C. D., Darby, D. A. and Hodgson, D. A. (2005) Late Pleistocene ice export events into the Arctic Ocean from the M'Clure Strait Ice Stream, Canadian Arctic Archipelago. *Global and Planetary Change*. 49: 139-162.

Stokes, C. R., Clark, C. D. and Winsborrow, M. C. M. (2006) Subglacial bedform evidence for a major palaeo-ice stream and its retreat phases in Amundsen Gulf, Canadian Arctic Archipelago. *Journal of Quaternary Science*. 21: 399-412.

Stokes, C. R., Clark, C. D., Lian, O. B. and Tulaczyk, S. (2007) Ice stream sticky spots: A review of their identification and influence beneath contemporary and palaeo-ice streams. *Earth-Science Reviews*. 81: 217-249.

Stokes, C. R., Lian, O. B., Tulaczyk, S. and Clark, C. D. (2008) Superimposition of ribbed moraines on a palaeo-ice-stream bed: implications for ice stream dynamics and shutdown. *Earth Surface Processes and Landforms*. 33: 593-609.

Stokes, C. R., Clark, C. D. and Storrar, R. (2009) Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet. *Quaternary Science Reviews*. 28: 721-738.

Stokes, C.R., Margold, M. Clark., C. D. Tarasov, L. (2016) Ice stream activity scaled to ice sheet volume during Laurentide Ice Sheet deglaciation. *Nature*. 530: 322-326.

Studinger, M., Bell, R. E., Blankenship, D. D., Finn, C. A., Arko, R. A., Morse, D. L. and Joughin, I. (2001) Subglacial sediments: A regional geological template for ice flow in West Antarctica. *Geophysical Research Letters*. 28: 3493-3496.

Swinthinbank, C. W. M. (1954) Ice streams. *Polar Record*. 7: 185–186.

Thomas, R. (1979) The Dynamics of marine ice sheets. *Journal of Glaciology*. 24: 167-177.

Truffer, M. and Echelmeyer, K. A. (2003) Of isbrae and ice streams. *Annals of Glaciology.* 36: 66-72

Tyrrell, J. B. (1887) Report on a part of Northern Alberta and portions of adjacent districts of Assiniboia and Saskatchewan. *Geological Survey of Canada.* 2: 150-152.

Van der Kamp, G. (2001) Methods for determining the in situ hydraulic conductivity of shallow aquitards – an overview. *Hydrogeology Journal*. 9: 5–16.

Vaughan, D. G., Smith, A. M., Corr, H. F. J., Jenkins, A., Bentley, C. R., Stenoienm, M. D., Jacobs, S. S., Kellog, T. B., Rignot, E. and Lucchitta, B. L. (2001) A review of Pine Island Glacier, West Antarctica: Hypotheses of instability vs. observations of change. In: Alley, R. B. and Bindschadler, R. A. (eds.) The West Antarctic Ice Sheet: Behaviour and Environment. *American Geophysical Union Antarctic Research Series*. 77: 257-281.

Vorren, T. O. and Laberg, J. S. (1997) Trough Mouth Fans – Palaeoclimate and Ice Sheet Monitors. *Quaternary Science Reviews*. 16: 865-881.

Wellner, J. S., Lowe, A. L., Shipp, S. S. and Anderson, J.B. (2001) Distribution of glacial geomorphic features on the Antarctic continental shelf and correlation with substrate. *Journal of Glaciology*. 47: 397-411.

Whillans, I. M., Bolzan, J. and Shabtaie, S. (1987) Velocity of Ice Streams B and C, Antarctica. *Journal of Geophysical Research*. 92: 8895-8902.

Whitaker, S. H. and Christiansen, E. A. (1972) The Empress Group in Saskatchewan. *Canadian Journal of Earth Science*. 9: 353-360.

Winsborrow, M. C. M., Clark, C. D. and Stokes, C. R. (2004) Ice Streams of the Laurentide Ice Sheet. *Géographie physique et Quaternaire*. 58: 269-280.

Appendix 1

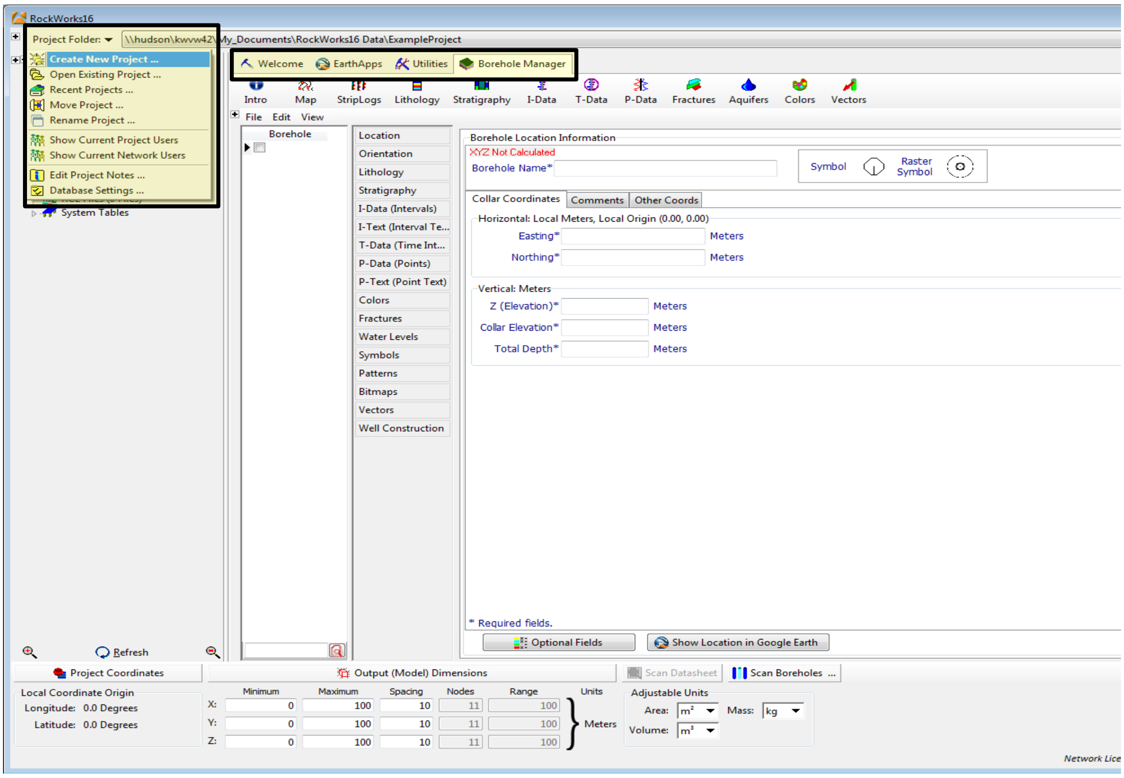
This appendix is intended to supplement the methods section of the thesis and contains a step-by-step guide to enable users to create lithologic and stratigraphic models using Rockworks 16™. This appendix also contains a supplementary file comprising a dynamic three-dimensional image of the static two-dimensional stratigraphic model included as a figure (Fig 5.4) in the main text. This diagram was created in Rockworks 3D 16™ modelling software package (Rockware Earth Science and GIS software: www.rockware.com) and is viewable and can be manipulated using the Rockworks 3D viewer Rockplot3D (available as a free software download at www.rockware.com/downloads/trailware.php [June 2016]).

Rockworks 3D viewer download location. Step by step introductory video is available and accessible from www.rockware.com/downloads/trailware.php

Creating a new project

When starting a project in Rockworks it is necessary to create a **New Project Folder**. To do this load the Rockworks 16™ program by clicking on the Rockworks shortcut on your desktop home screen. Two program screens will open a project screen and a help screen. Close the help screen. The project screen is the main screen base from here you will save, manipulate and input and export data. On the left of this screen you will see all files created. Within the centre of the screen several parts of the program can be accessed including a **Welcome**, **EarthApps, Utilities and Borehole Manager** screen. Familiarise yourself with this screen.

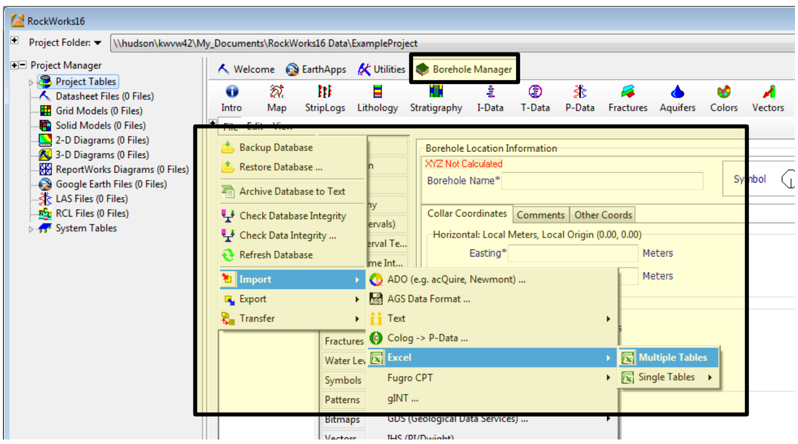
To create a new project click on the **Borehole Manager** tab and choose project **Folder → Create New Project** to display the **Create New Project** dialog box. Click on the yellow folder icon to specify where you would like your new project to be saved. Click on the **Next** button to accept the default settings and proceed to the **Project Coordinates** step. Here select the **Coordinate System** you wish to work in and verify the **Datum** if necessary. Select the **Vertical Units** you wish to work in. Click the **Yes** button to create your new project.



Rockworks 16™ home screen. Highlighted are the: ***Welcome, EarthApps, Utilities*** *and* ***Borehole Manager*** *tabs. The dropdown Project folder tab is also expanded to display the location of the* ***Create New Project*** *button.*

<u>Data Compilation- Importing and synthesising borehole data</u>

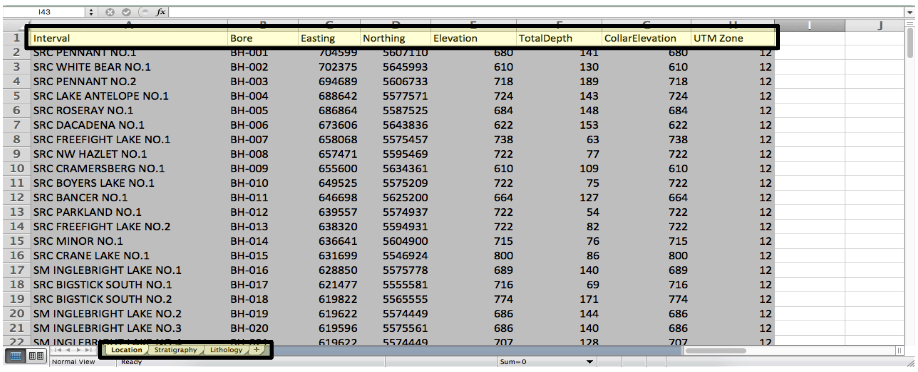
There are several methods to import data into Rockworks. The easiest and quickest method, especially in cases where there is a lot of data, is via an Excel file. Data corresponding to Rockworks can be stored on multiple sheets within the Excel File. Note that in order for Rockworks to automatically recognise the columns in Excel you should only use those defined by Rockworks. If you would like to add additional names Rockworks will still add this data however they will also have to be manually added to Rockworks.



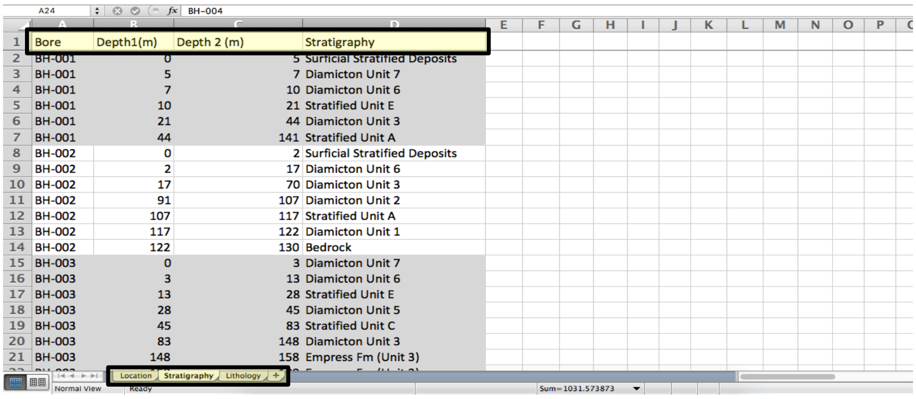
Rockworks home screen. Within the Borehole Manager the ***File→ Import→ Excel→ Multiple Tables*** *drop down menus highlighted.*

Once you have created an Excel spreadsheet, go back and open Rockworks. Within Rockworks, choose the **File → Import → Excel → Multiple Tables** (or **Single Table** if you only have a single sheet table) menu commands. The **Confirm** window will then appear that prompts you to **Backup** your database. While you may not want to do this now, it is best to get into the habit of backing up the system prior to importing new data in case unwanted changes occur.

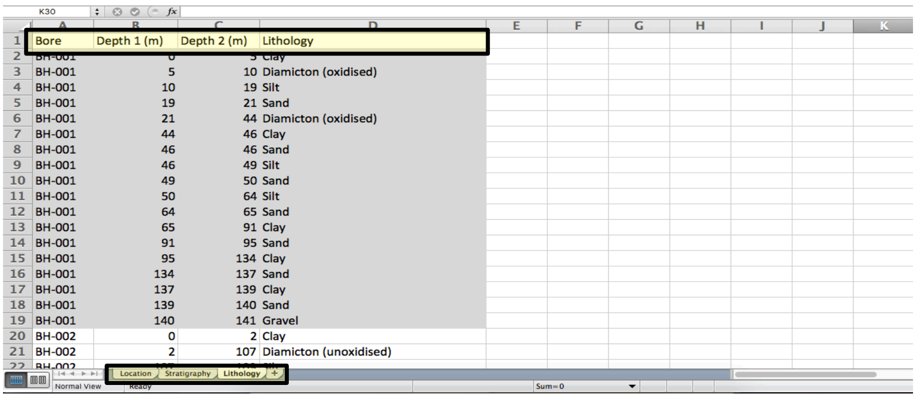
Note when creating your Excel datasheet boreholes IDs should contain the same amount of numerial figures. For example for data sets with +99 boreholes, borehole IDs should be numberd BH-001, BH-002, BH-003... rather than BH-1, BH-2, BH-3, ensuring boreholes are displayed in the correct order in the Borehole Manager.



	Interval	Bore	Easting	Northing	Elevation	TotalDepth	CollarElevation	UTM Zone
2	SRC PENNANT NO.1	BH-001	704599	5607110	680	141	680	12
3	SRC WHITE BEAR NO.1	BH-002	702375	5645993	610	130	610	12
4	SRC PENNANT NO.2	BH-003	694689	5606733	718	189	718	12
5	SRC LAKE ANTELOPE NO.1	BH-004	688642	5577571	724	143	724	12
6	SRC ROSERAY NO.1	BH-005	686864	5587525	684	148	684	12
7	SRC DACADENA NO.1	BH-006	673606	5643836	622	153	622	12
8	SRC FREEFIGHT LAKE NO.1	BH-007	658068	5575457	738	63	738	12
9	SRC NW HAZLET NO.1	BH-008	657471	5595469	722	77	722	12
10	SRC CRAMERSBERG NO.1	BH-009	655600	5634361	610	109	610	12
11	SRC BOYERS LAKE NO.1	BH-010	649525	5575209	722	75	722	12
12	SRC BANCER NO.1	BH-011	646698	5625200	664	127	664	12
13	SRC PARKLAND NO.1	BH-012	639557	5574937	722	54	722	12
14	SRC FREEFIGHT LAKE NO.2	BH-013	638320	5594931	722	82	722	12
15	SRC MINOR NO.1	BH-014	636641	5604900	715	76	715	12
16	SRC CRANE LAKE NO.1	BH-015	631699	5546924	800	86	800	12
17	SM INGLEBRIGHT LAKE NO.1	BH-016	628850	5575778	689	140	689	12
18	SRC BIGSTICK SOUTH NO.1	BH-017	621477	5555581	716	69	716	12
19	SRC BIGSTICK SOUTH NO.2	BH-018	619822	5565555	774	171	774	12
20	SM INGLEBRIGHT LAKE NO.2	BH-019	619622	5574449	686	144	686	12
21	SM INGLEBRIGHT LAKE NO.3	BH-020	619596	5575561	686	140	686	12
22	SM INGLEBR[illegible]	[illegible]	619622	5574449	707	128	707	12

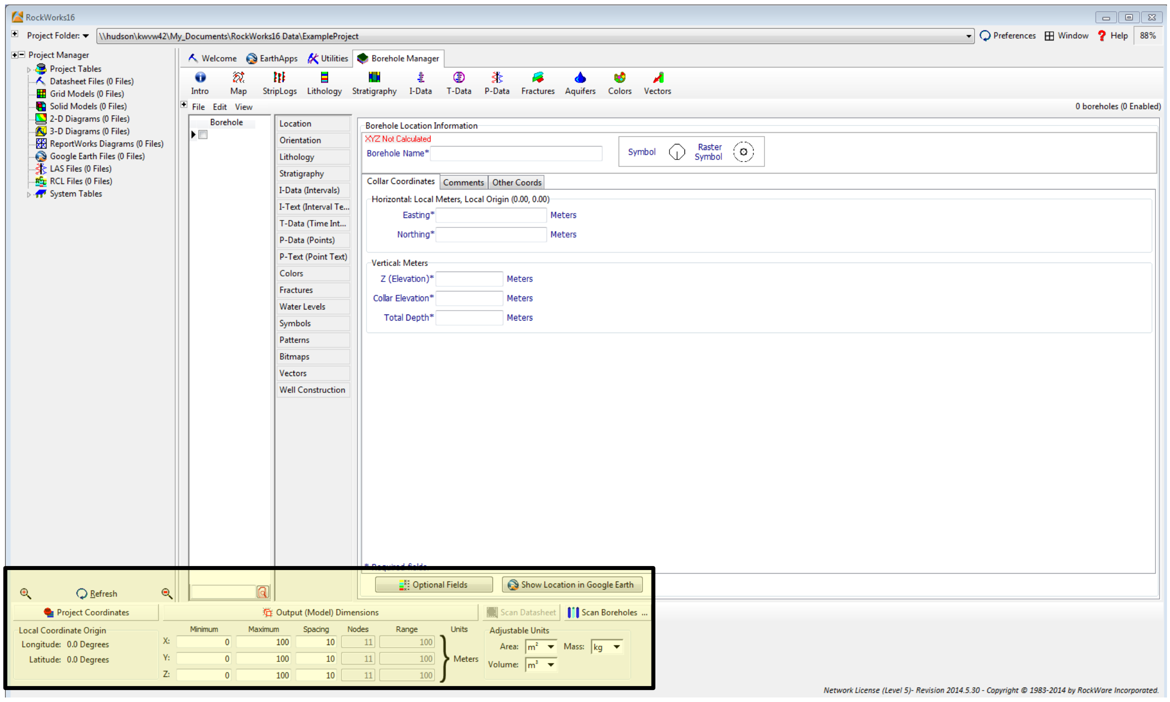


	Bore	Depth1(m)	Depth 2 (m)	Stratigraphy
2	BH-001	0	5	Surficial Stratified Deposits
3	BH-001	5	7	Diamicton Unit 7
4	BH-001	7	10	Diamicton Unit 6
5	BH-001	10	21	Stratified Unit E
6	BH-001	21	44	Diamicton Unit 3
7	BH-001	44	141	Stratified Unit A
8	BH-002	0	2	Surficial Stratified Deposits
9	BH-002	2	17	Diamicton Unit 6
10	BH-002	17	70	Diamicton Unit 3
11	BH-002	91	107	Diamicton Unit 2
12	BH-002	107	117	Stratified Unit A
13	BH-002	117	122	Diamicton Unit 1
14	BH-002	122	130	Bedrock
15	BH-003	0	3	Diamicton Unit 7
16	BH-003	3	13	Diamicton Unit 6
17	BH-003	13	28	Stratified Unit E
18	BH-003	28	45	Diamicton Unit 5
19	BH-003	45	83	Stratified Unit C
20	BH-003	83	148	Diamicton Unit 3
21	BH-003	148	158	Empress Fm (Unit 3)



	Bore	Depth 1 (m)	Depth 2 (m)	Lithology
2	BH-001	0	5	Clay
3	BH-001	5	10	Diamicton (oxidised)
4	BH-001	10	19	Silt
5	BH-001	19	21	Sand
6	BH-001	21	44	Diamicton (oxidised)
7	BH-001	44	46	Clay
8	BH-001	46	46	Sand
9	BH-001	46	49	Silt
10	BH-001	49	50	Sand
11	BH-001	50	64	Silt
12	BH-001	64	65	Sand
13	BH-001	65	91	Clay
14	BH-001	91	95	Sand
15	BH-001	95	134	Clay
16	BH-001	134	137	Sand
17	BH-001	137	139	Clay
18	BH-001	139	140	Sand
19	BH-001	140	141	Gravel
20	BH-002	0	2	Clay
21	BH-002	2	107	Diamicton (unoxidised)

Example of standardised Rockworks compatible Excel data tables. Column names used here are recognised automatically by Rockworks 16™.



Project dimension setting tool. Users should reset the dimensions of the project when they add or alter borehole input data.

To select the file, click on the Yellow **Open Folder** icon and locate your file. Click **Next** and a **Block Selection** tab should appear. The **Block Selection** tab is used to match the names of sheets in Excel with **Database Table Names** in Rockworks. If you have used names that do not match up automatically this is the point to assign them manually. Click **Next**. The **Data Coordinates** tab should appear. This tab lets you specify the **Coordinate System** of the incoming data. Set this to the correct **Coordinate System**. Click **Next**. The **Location** tab will now appear. This tab lets you specify the columns in the Excel spreadsheet that correspond with the field in the Rockworks **Location Table**. At the side of the window, borehole overwrite options are available. These options allow you to alter how data is imported. To import a new record select **Create New Record**. Click **Finish** and **Confirm**. Rockworks will now load the data and calculate the XY location and elevation of each data point within the borehole record you have imported. Note: importing of large datasets may take some time. Click **Yes** to display the output units and dimensions.

Before you start to create maps and models within Rockworks you must set the project dimensions. To do this select **Scan Boreholes** at the bottom of the home screen. Click **Process**.

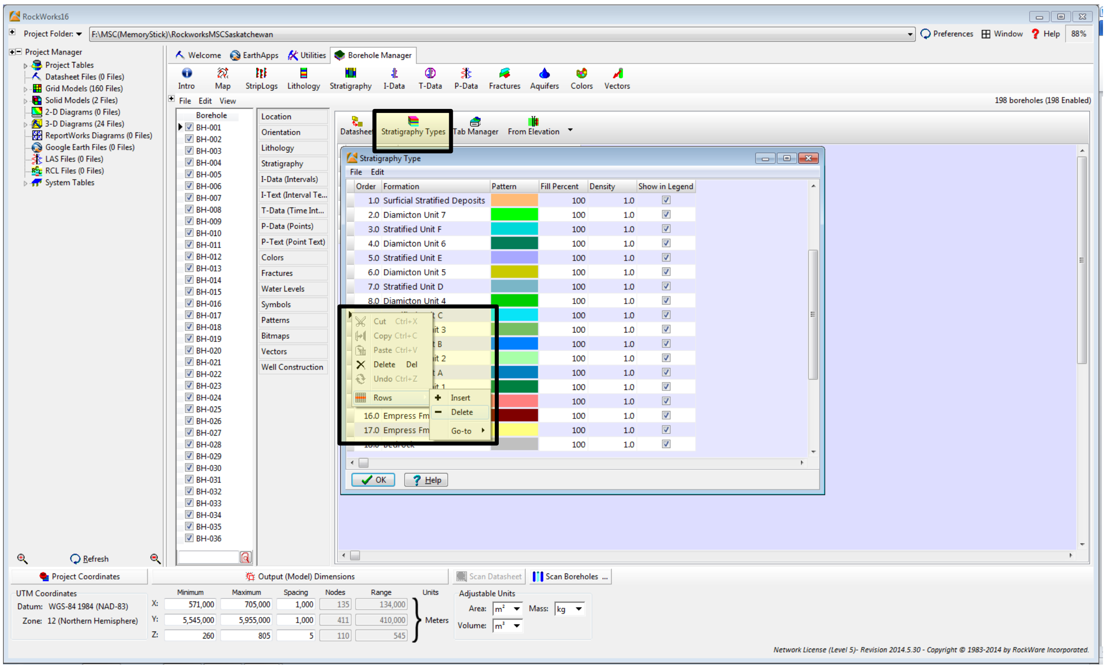
<u>Visualising Borehole Distribution-Site Map construction</u>
In order to interpret the stratigraphy of your study area, it is necessary to view data stored in the **Borehole Manager** using striplogs, cross-sections and other tools. However prior to this, visualisation of the spatial distribution of your boreholes sites must be completed.

Within the borehole manager select your project folder. Within this select **Map → Borehole Locations**. This will open the **2-Dimensional Borehole Location Map** window. Within this window adjust settings to your desired preferences (Click on **Borehole Symbol + Label Options** for options to change symbols and the size of borehole ID text). Click on the **Process** button.

<u>Altering lithologic and stratigraphic type tables</u>
Prior to modelling stratigraphy and lithology it is necessary to set the lithology and stratigraphy type tables. The types tables define the colour and pattern of lithology and stratigraphy. To do this click on the **Borehole Manager** tab on the home screen and select **→ Borehole 001 → Lithology (Stratigraphy)**. This will open the **Lithology (Stratigraphy) Type Table**. Within this table will be a list of all lithology (stratigraphy) types that you recorded within Excel. You may now manually alter the colours and patterns of units and or import a standardised set of colours and patterns from another project via the **File → Inset** button.

Note if any naming inconsistencies occurred when you input your data in Excel, multiple keywords for the same material will now appear. To change this, right click on the material and click **Delete**. An option will now appear to transfer these layers to the correctly spelled material section. Click.

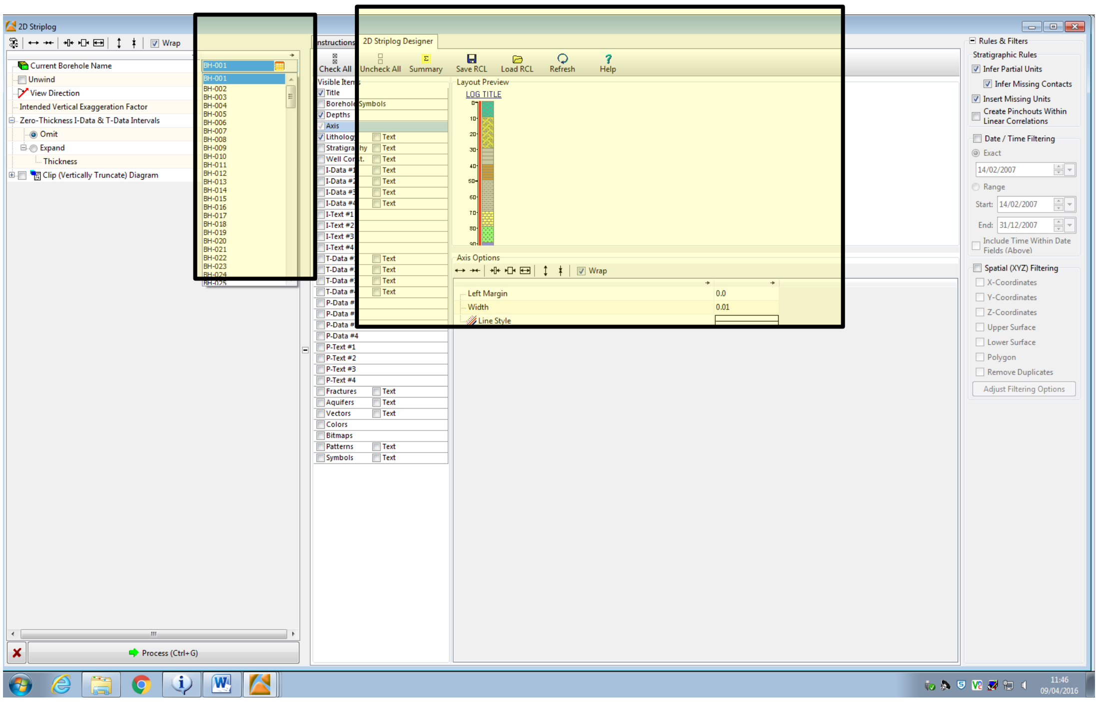
It is also necessary with the stratigraphy table to order the units in terms of stratigraphic position. From 1 (lowest unit) to 100+ (highest unit). If this process is not completed stratigraphy modelling will produce unrealistic models.



Stratigraphy type table associated with stratigraphic units within this study. The delete option used to remove an unwanted stratigraphic unit is highlighted.

<u>Cross Lithologic section and single log construction</u>

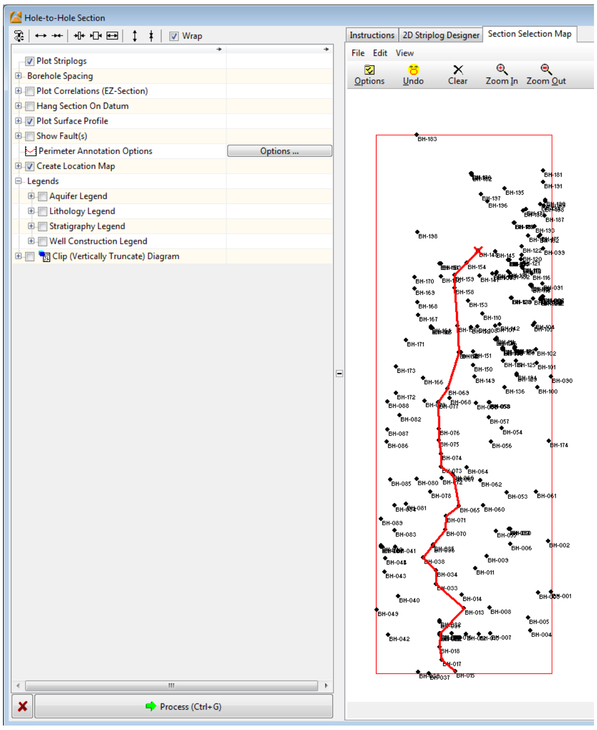
In many cases it is useful to create a **Single Striplog** of a single borehole. This is often useful to aid interpretation of stratigraphy or to check previously interpreted units. To do this click on the **Borehole Manager** (home screen) and select the desired core. Go to the **Striplogs Menu** and choose **Striplogs → 2-Dimensional → Single log.** A **2D Striplog** option window will now open. Click on the **2D Striplog Designer**. This Window allows you to establish what data will be displayed in the striplog. Select any other desired options. Click **Process.**



Single striplog designer used to create single composite striplogs. The borehole selection tool and dropdown menu are highlighted on the left. The 2D striplog designer tab has been selected; this tab is used to change the order and design of specific sections of a log (highlighted to the right).

Creating a lithologic striplog section

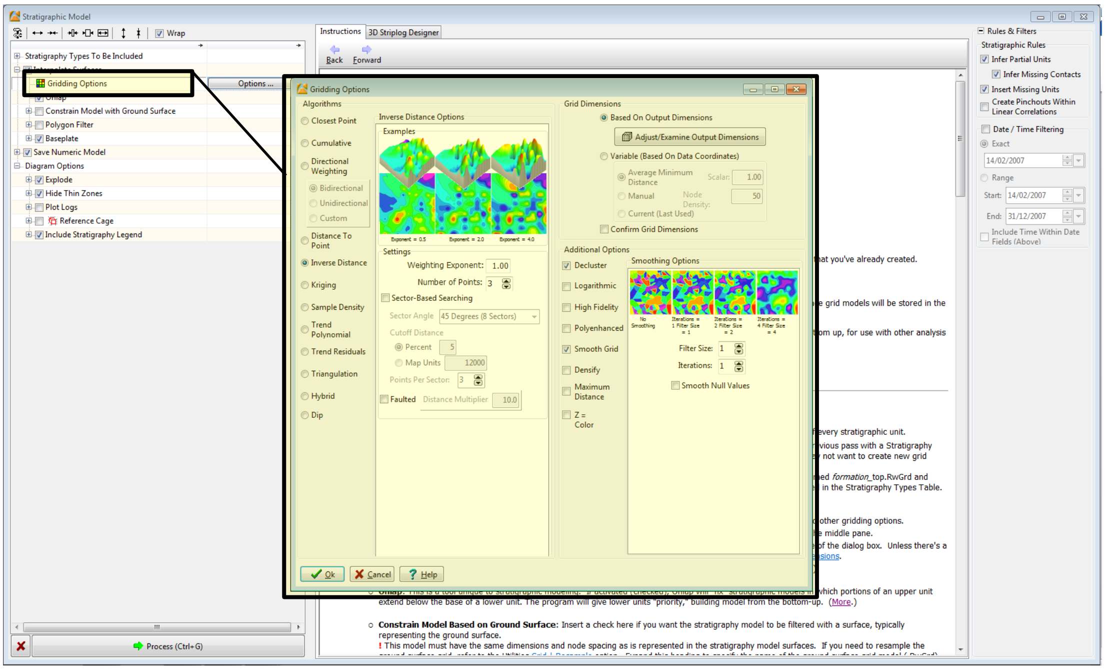
This is a useful tool to view the lithology of a section of your study area. To create a striplog section Choose the **Striplogs → 2-Dimensional → Section Menu** command to display the **Hole-to-Hole Section window**. Click the **2D striplog** designer tab to display striplog options. Select your desired options. Next, click on the **Section Selection Map** (to the right of the 2D Striplog Designer tab) and select your desired section. Click **Process.**

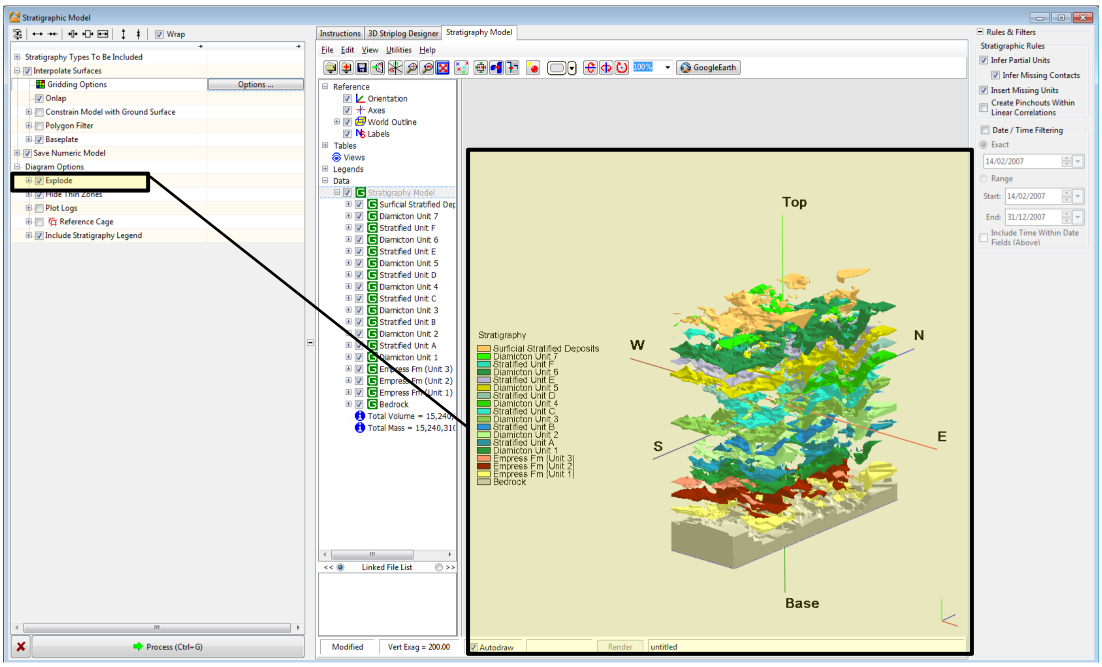


Striplog section tool showing section selection map. Within the section selection map all enabled boreholes are shown and can be selected to form part of a striplog section.

Creating 3D stratigraphic model and diagrams.

Once you have interpreted all stratigraphic units, either within Excel or through a combination of manual interpretation and cross referencing using lithologic sections and logs discussed above you can now create a **Stratigraphic Model.** It is likely that once you have create a stratigraphic model in 3D, your interpretations may change and you may have to repeat several of the above steps in order to create your final outputs. The simplest method of creating stratigraphic modelling is described below. However models can also be created manually and imported and appended. To create a simple stratigraphic model: go to the stratigraphy menu and choose **Stratigraphy → Model**. Add a check to the **Interpolate Surfaces** box. Expand the interpolate options by clicking the **+** sign. Click on the **Options** button to access various modelling settings. Next click on the **Algorithms** section. Here you will see a list and descriptions of different interpolation methods. Select the appropriate method for your data. Click the **OK** button to close the **Gridding Options** window. Click **Process.** A stratigraphic model will now be generated. To change settings on this model alter options in the left hand panel and click process. Note if your diagrams have multiple stratigraphic layers you may want to select the **Explode** button to create an offset 'exploded diagram' (see main thesis text for examples).

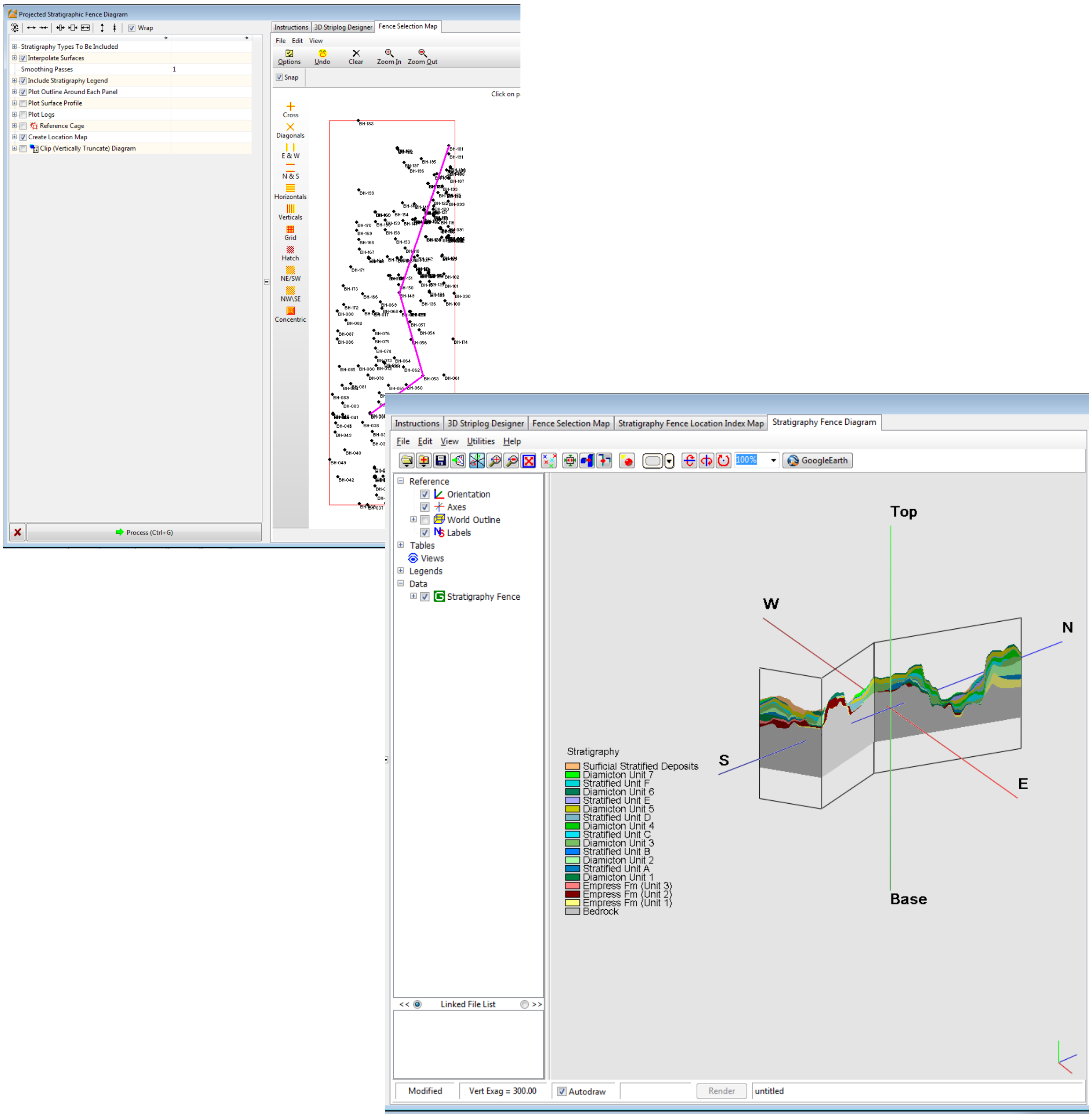




Stratigraphic model creation window and resulting 3D model window. The gridding option tab and window is highlighted and shows the options used within this study. The function of the explode tool is also highlighted as well as the resulting exploded stratigraphic model.

Creating stratigraphic fence diagrams

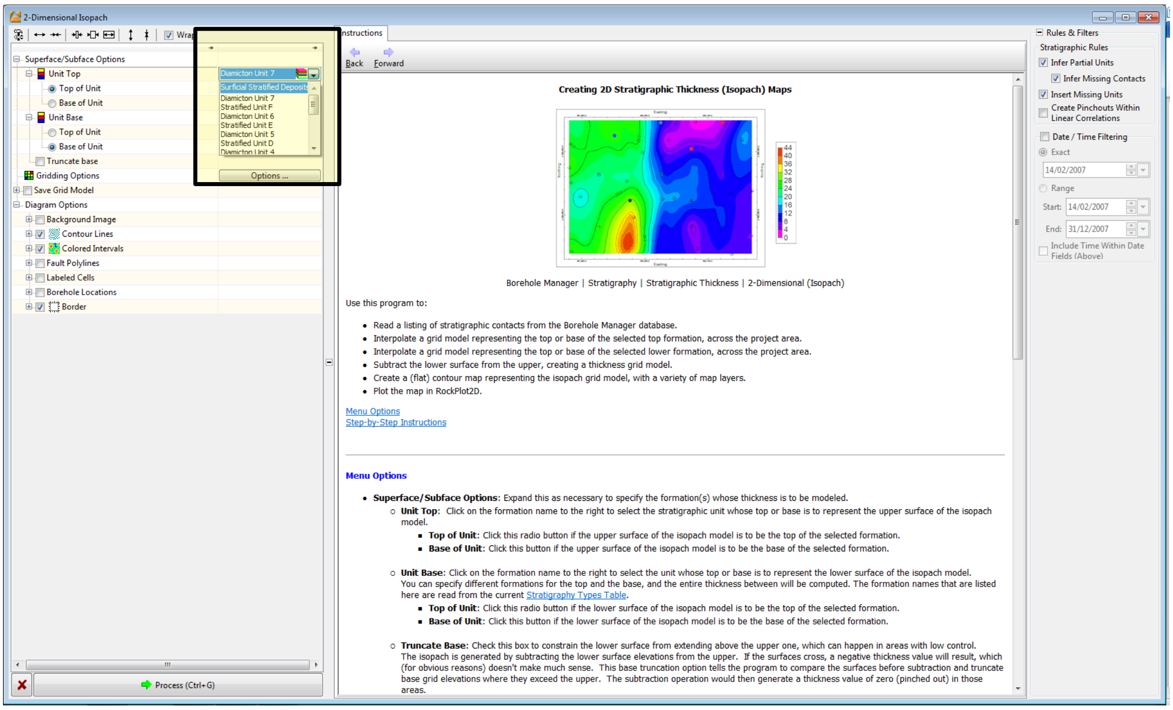
This is a useful tool to allow you to view a variety of angles in your stratigraphic model, which may aid further interpretations. To build a stratigraphy fence diagram go to **Stratigraphy → Fence (3D profiles) → Fence Diagrams**. Click the **Fence Selection Map** tab to the right. You should now see a map of all your boreholes. Select an appropriate fence. Click **Process**. A new **Rockplot 3D** tab will open with a modelled fence diagram displayed.



Fence diagram creation window. Fence selection map and resulting fence diagram are shown. Note several fence diagrams can be created simultaneously to provide a more comprehensive view of a users project.

2D isopach creation

This is a useful step that allows you to view each stratigraphic unit in 2D. This may be a step you wish to take between running stratigraphic models to aid in your interpretations. Within the borehole manager home screen choose **Stratigraphy → Structural Elevation → 2-D**. Select your desired stratigraphy unit. Click on the **Grid (Output)** and name the grid. Under **Gridding Options** you have the chance to change these from the original options used in the stratigraphic model. Click on the **+** sign to alter these. When you have made your desired changes click **Process**. A **Rockplot2D Structure Map** should now appear. If a legend is not already displayed you can insert one using the option on the top of the map window.



***2D Isopach** creation window. The stratigraphic unit selection dropdown menu is highlighted.*

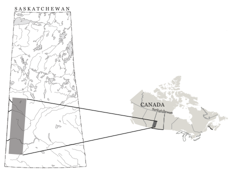
Document export and manipulation

Once models, cross sections, and or logs have been created to the users preference, files can be exported in **DXF, BMP, JPG, TIFF, PNG, ESRI** shapefiles. It is also worth noting that the majority of graphics can be exported to Google Earth. This feature is particularly useful for viewing single borehole point data in relation to geomorphology. Once data has been exported, files can also be manipulated in a graphics package such as Adobe Illustrator. This proves particularly useful for manipulating 2D diagrams and or overlying multiple diagrams to produce 3D conceptual diagrams. Furthermore multiple models can also be **Appended, Draped** or **Overlain** over one another within Rockworks. These functions are particularly useful when the relationship between stratigraphic units and topographic features needs to be assessed.

GLACIAL GEOMORPHOLOGY and SURFICIAL GEOLOGY of the SOUTHWESTERN SASKATCHEWAN SWATH

Map Sheet 1

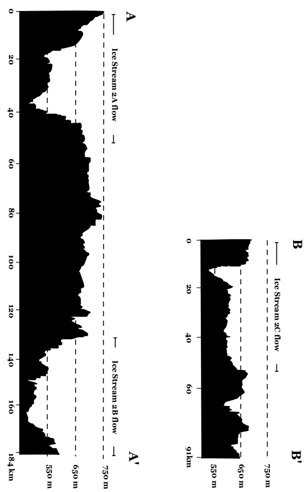
(NTS 73F, 73C 72N, 72K)



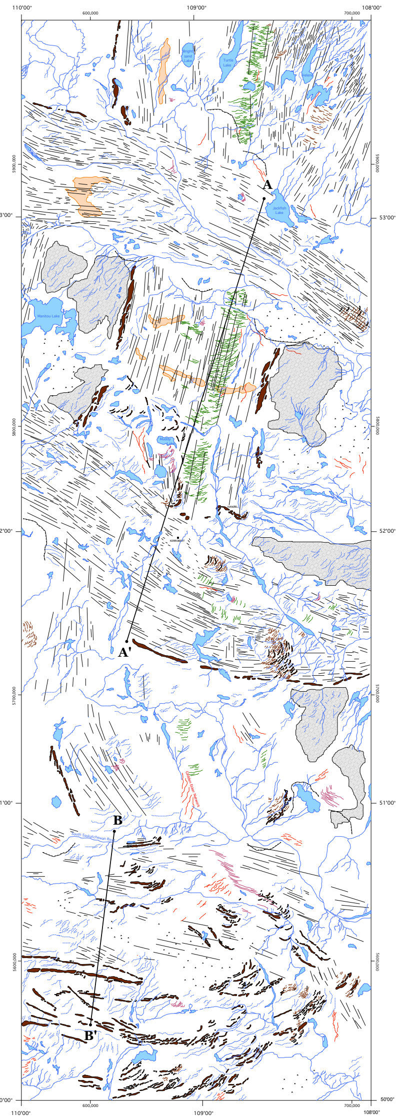
Kilometers 5 0 10 20 30 40 50 60 70 80 Kilometers

Miles 5 0 10 20 30 40 50 60 Miles

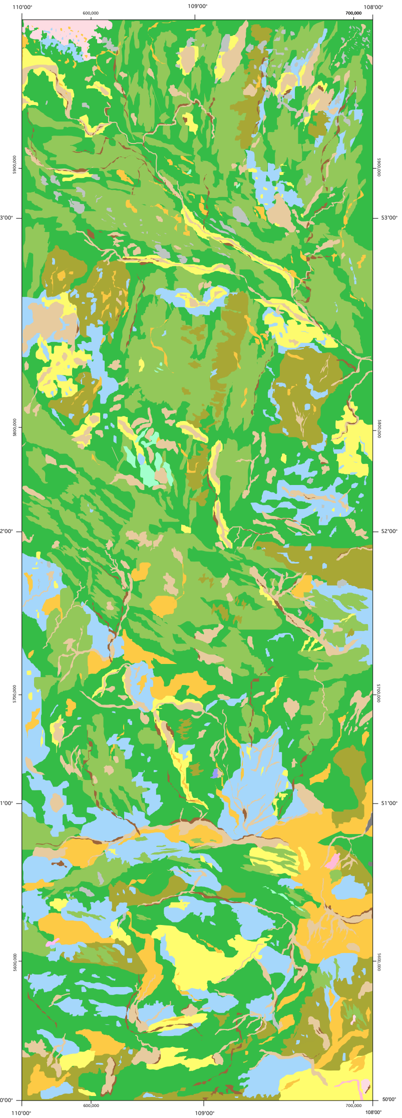
Projection: UTM 12U
Datum: North American Datum 1983



Transects of topography



Glacial geomorphology of the southwestern Saskatchewan swath



Surficial geology of the southwestern Saskatchewan swath. Adapted from: Campbell (1986a, b, 1987a, b)

LEGEND

Rivers- *Active channels*

Ponds and lakes

Dunes- *Semi stablised sand dunes and blowouts*

Meltwater channels- *Channels cut by former glacial spillways*

Hummocky terrain- *Hummocky, dead-ice and disintergration moraine; includes prairie mounds and some transversely lineated moraines and local pitted laustrine deposits*

Moraine crest (major)- *Ridges greater than 50 m to 200 m wide; interlobate, ice stream lateral and marginal moraines*

Moraine crest (minor)- *Ridges less than 20 m wide and usually less than 2 m high includes interlobate, ice stream lateral and marginal moraines*

Overridden moraine crest- *Moraine ridges exhibiting geomorphological evidence for ice overriding*

Ice flow parallel lineations- *Includes glacial fluting, drumlinoid ridges, drumlins and mega-scale glacial lineations*

Eskers, kame, kame-complex

Escarpment- *Bedrock, ice contact*

Geometric ridge networks- *Comprising till cores, straight or slightly arcuate, and intersecting ridges*

Ice-thrust ridges- *50-1500 m parallel sharply-crested ridges*

Organic deposits- *Black, organic rich peat; commonly occurs as blankets found in bogs occurring in low-lying, poorly drained areas*

Scree deposits- *Poorly sorted, sandy to gravelly material; found along slopes and at the base of steep bedrock cliffs*

Fluvial deposits- *Sand and gravels deposited by a modern stream; commonly occur as terraces and floodplains in river valleys; moderately to well stratified and sorted*

Lacustrine deposits- *Sediment deposited in and adjacent to lakes; includes beach deposits. Predominantly sand and gravel, stratifed to massive, generally moderately to well sorted*

Aeolian deposits- *Wind-deposited sediment; medium to fine grained sand; well sorted; generally massive; local cross bedding or ripple laminations*

Glaciolacustrine deposits- *Sediment deposited in glacial meltwater lakes; sand, silt and minor clay and gravel; generally well stratifed and sorted*

Glaciofluvial deposits- *Fine to coarse sand and pebble to cobble-sized gravels; deposits are typically found in terraces and outwash plains and in association with meltwater channels*

Morainic deposits- *Terrain consisting of unstratifed, unsorted sediment deposited by a glacier; mainly till (a mixture of sand, silt and clay); locally may be composed of one or more of shale, siltstone, sandstone or stratfied drift, or includes discontinuous layers of stratifed sediment generally sand*

Fluted morainic deposits- *All glacially streamlined terrain; varies from alternating furrows and ridges to nearly equidimensional smoothed hills; all landforms parallel to the local glacier flow direction. Includes flutes, drumlins and mega-scale glacial lineations; till with local sand and gravel*

Stagnant ice morainic deposits- *Terrain resulting from the collapse and lateral movement of supraglacial sediment in response to the melting of buried stagnant ice*

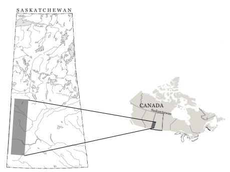
Ice-thrust morainic deposits- *Masses of originally subglacial sediment incorporated, transported and deposited by a glacier more or less intact; deposits may include syngenetic till and masses of pre-existing till, stratifed drift and/ or bedrock*

Bedrock deposits

SEDIMENT THICKNESS, ICE STREAM LOCATIONS and BEDROCK TOPOGRAPHY of the SOUTHWESTERN SASKATCHEWAN SWATH

Map Sheet 2

(NTS 73F, 73C 72N, 72K)



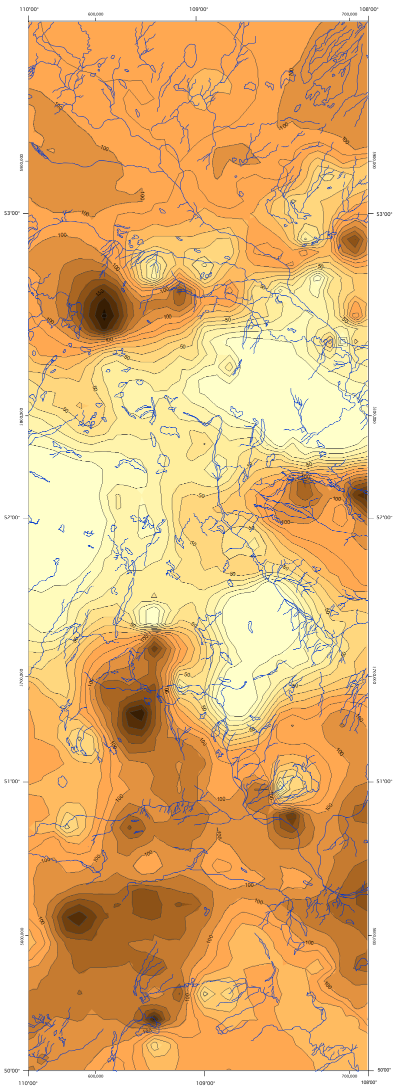
Kilometers 5 0 10 20 30 40 50 60 70 80 Kilometers

Miles 5 0 10 20 30 40 50 60 Miles

Projection: UTM 12U
Datum: North American Datum 1983

Sediment Thickness (m)

- 170-180
- 160-170
- 150-160
- 140-150
- 130-140
- 120-130
- 110-120
- 100-110
- 90-100
- 80-90
- 70-80
- 60-70
- 50-60
- 40-50
- 30-40
- 20-30
- 10-20
- 0-10

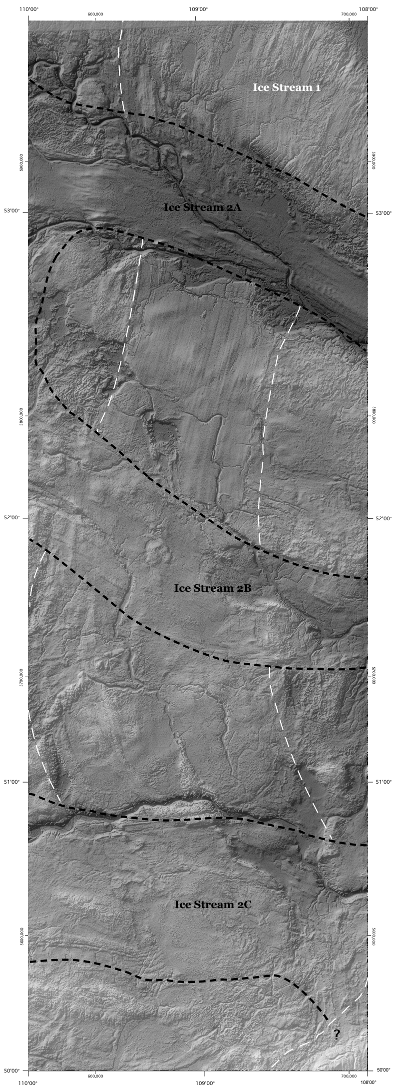


Sediment thickness of the southwestern Saskatchewan swath

Elevation (m a.s.l.)

High: 942

Low: 460

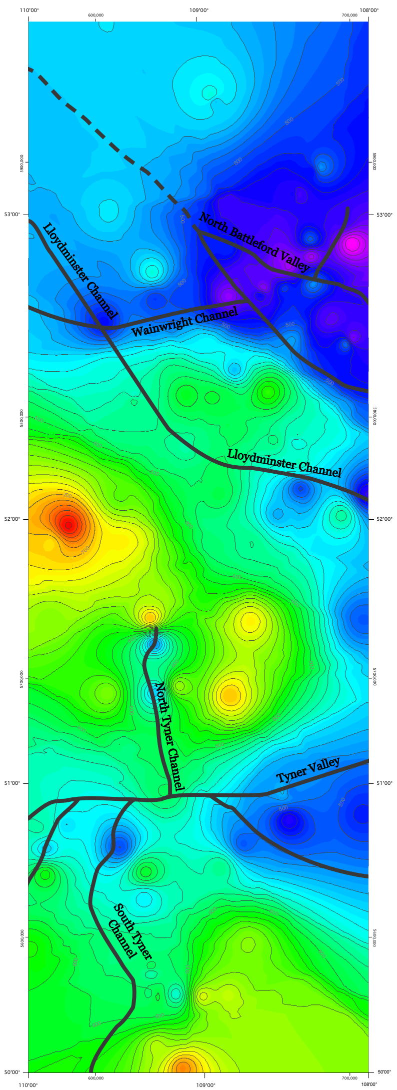


Shuttle Radar Topography Mission (SRTM) image of ice stream flow paths in the southwestern Saskatchewan swath

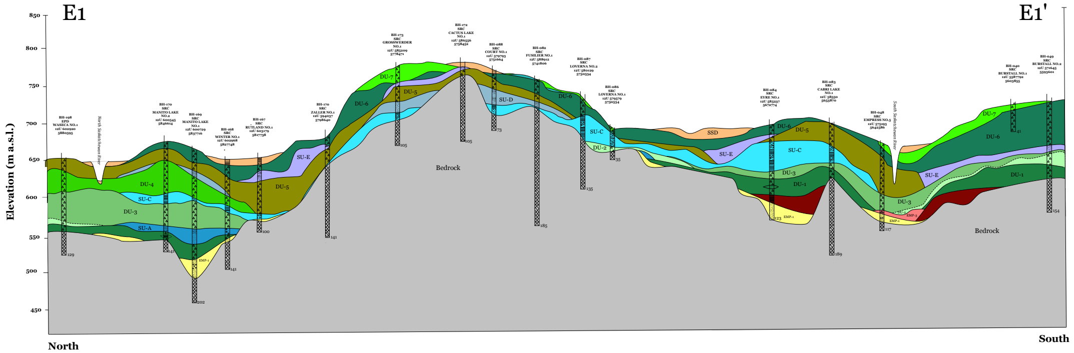
Elevation (m a.s.l.)

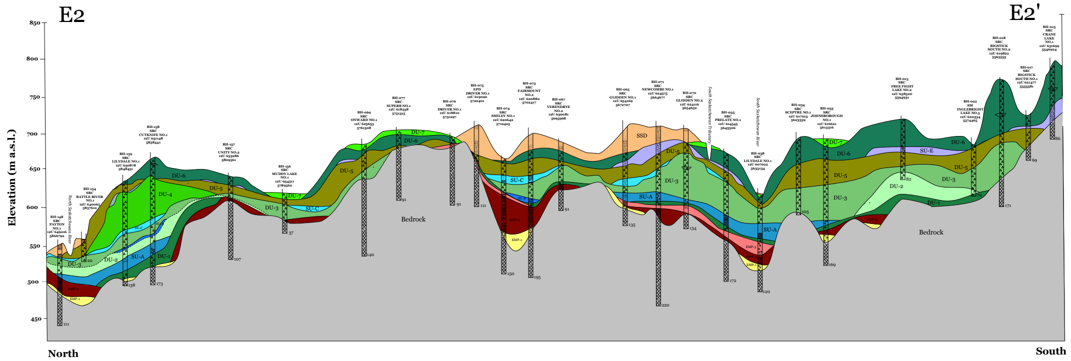
High: 760

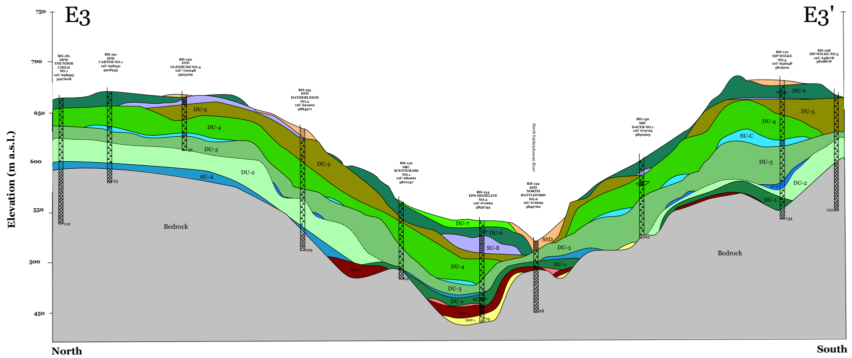
Low: 400



Bedrock topography of the southwestern Saskatchewan swath. Solid black lines denote buried valleys adapted from: Stalker (1961). Dashed line represents potential buried valleys



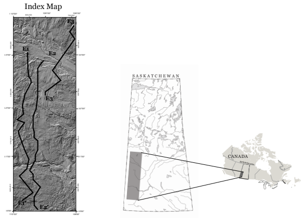


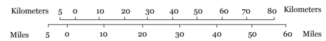


STRATIGRAPHIC UNITS

Surfical Stratified Deposits (SSD)	*Stratified gravel, sand, silt and clay of fluvial, lacustrine or aeolian origin*
Diamicton Unit 7 (DU-7)	*Glacial diamicton, silty-sandy*
Stratified Unit F (SU-F)	*Stratified sand, gravel and silt of glaciofluvial or glaciolacustrine origin*
Diamicton Unit 6 (DU-6)	*Clayey and sandy glacial diamicton contains incorporated masses of glacially displaced sediment*
Stratified Unit E (SU-E)	*Stratified sand and gravel and small amounts of silt and clay of glaciolacustrine or glaciofluvial origin*
Diamicton Unit 5 (DU-5)	*Silty-sand rich glacial diamicton; very coarse sand rich in carbonate fragments*
Stratified Unit D (SU-D)	*Stratifed sand and gravel of glaciofluvial origin*
Diamicton Unit 4 (DU-4)	*Clayey glacial diamicton*
Stratified Unit C (SU-C)	*Stratified silt and clay in many locations also contains small amounts of sand and gravel of glaciolaustrine origin*
Diamicton Unit 3 (DU-3)	*Sand rich glacial diamicton*
Stratified Unit B (SU-B)	*Stratified clay, sand and gravel of glaciofluvial origin*
Diamicton Unit 2 (DU-2)	*Clay rich glacial diamicton*
Stratified Unit A (SU-A)	*Silt, sand and gravel of glaciofluvial origin*
Diamicton Unit 1 (DU-1)	*Clayey glacial diamicton interbedded with well sorted clays silts and sands, overlain by stratified sediment in some places*
Empress Group (EMP-3)	*Stratified sand and gravel of glaciofluvial origin*
Empress Group (EMP-2)	*Stratified silt and clay of lacustrine or fluvial origin*
Empress Group (EMP-1)	*Stratified sand and gravel, mainly chert and quartzite derived from the Cordilleran Mountains. Preglacial fluvial origin*
Bedrock	*Defined as all lithified clastic sediment that underlies the early Tertiary erosional surface. Commonly of crystalline Precambrian rock overlain by sedimentary rock cover*

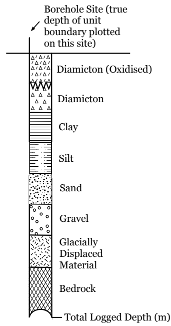
Index Map





Projection: UTM 12U
Datum: North American Datum 1983
Vertical Exaggeration (E1-E1'/E2-E2' 180 x)
(E3-E3' 210 x)

Borehole Key



Boreholes may show various combinations of lithologies shown above

QUATERNARY STRATIGRAPHY of the SOUTHWESTERN SASKATCHEWAN SWATH

Map Sheet 3

(NTS 73F, 73C 72N, 72K)